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REFERENCE

Lane, B. H., 1935, Suggestions to authors of papers submitted for publication by the United States Geological Survey: U. S. Geol. Survey, Washington, D. C. (For sale by Superintendent of Documents, Washington, D. C. 35c.)

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MARCH 1955

QUARTZITE PEBBLE DEFORMATION IN CENTRAL VERMONT

WILLIAM F. BRACE

ABSTRACT. Quartzite pebbles from thin conglomerate beds in central Vermont are parallel to bedding and elongate in a direction subnormal to fold axes. Pebble form is approximately ellipsoidal with a greatest to least axis ratio ranging up to 20.

Grain dimensional orientation of the pebbles is weak and parallel to pebble elongation. No correlation is possible between degree of pebble elongation and the development of microscopic features such as lamellae, undulose extinction and fracturing.

Quartz axis orientation in the pebbles is strong and consists of maxima at positions VIII, IV and II and incomplete girdles intersecting in a line parallel to the intermediate axis of the pebble.

Independent field investigation and petrofabric analysis of the pebbles suggest that flattening parallel to bedding was the dominant process in pebble deformation. Elongation has been produced by slip on pairs of surfaces inclined to bedding.

Translation gliding may have been active in permitting slip and ultimate elongation of quartz grains. This hypothetical gliding system is based on [m:r] as the glide line in planes {m}, {r}, {c} and other unknown surfaces in the zone about [m:r]. Lack of symmetry and nonequivalence of axis concentrations may be due to the influence of fabric preserved from earlier stages in the history of the quartzite fragments.

INTRODUCTION

Fabric units such as stretched pebbles, fossils and ooids are particularly significant in the study of rock deformation inasmuch as initial and final shapes of the deformed materials are rather accurately known.

Following an areal study of a metamorphic complex in central Vermont (Brace, 1953), the writer selected deformed quartzite pebbles for special study. Heretofore in the Vermont area attention has been directed largely to analysis of pebble geometry and the relationship of pebble elongation to the minor structures in surrounding argillaceous rocks (Thompson, 1950, p. 104; Osberg, 1952, p. 30). For the present study petrofabric analyses of deformed pebbles were made in an attempt to correlate grain orientation with pebble shape, and to use these relationships to investigate the mechanics of quartz deformation.

Data on pebble geometry were obtained from 50 pebbles from seven conglomerate localities (Brace, 1953, p. 30, 46). As most of the pebbles are embedded in quartzite matrix, time did not permit removal and accurate shape measurement except for a few samples. Maximum and minimum pebble dimensions are the principal data obtained. Nine quartzite pebbles were sectioned and their lattice orientation obtained, using standard Universal Stage procedures as described by Fairbairn (1949, p. 241-290).

GEOLOGIC SETTING

An excellent recent summary of the regional geology of southern Vermont, southern New Hampshire and east-central New York may be found in the Geological Society of America Field Trip Guidebook of 1952 (Billings et al., 1952). Detailed descriptions of quadrangles in central Vermont have been published by Osberg (1952) and Brace (1953).

The Green Mountains of central Vermont are underlain by metamorphosed and deformed sedimentary and volcanic rocks ranging in age from Precambrian to Middle Ordovician (Billings et al., 1952, p. 14-23). Major unconformities occur near the base of the Paleozoic and within Middle Ordovician strata. The lowest member of sequences believed to be Paleozoic in age consists of graywacke, quartzite and thin lenses of conglomerate. These conglomerates, consisting in large measure of quartzite fragments embedded in a quartzite or graywacke matrix, provide the stretched pebbles of this study.

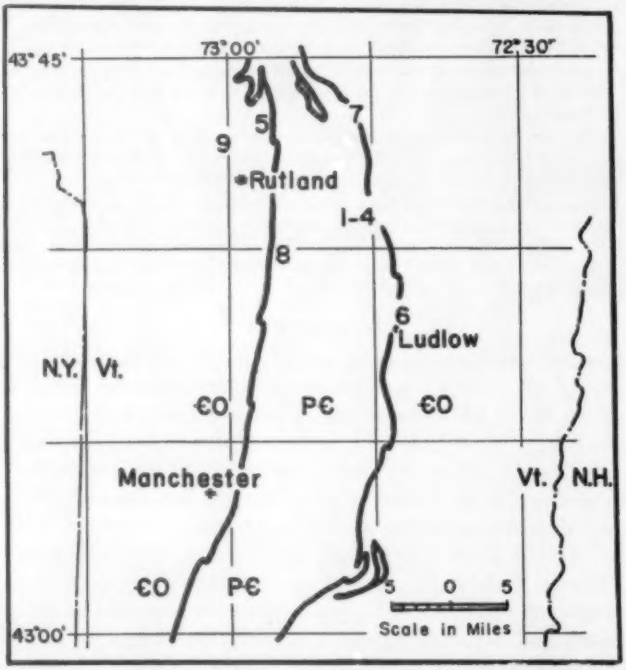


Fig. 1. Index map of Central Vermont showing outline of Precambrian basement complex and the conglomerate localities of this study (numbers 1 through 9). Sketched from Billings, et al., 1952.

The structural picture in central Vermont is dominated by a large anticlinorium trending about north-south. Precambrian rocks are exposed in the center of this large fold, as shown in figure 1. Paleozoic beds have been arched into folds overturned to the west, and thrust-faulted along east-dipping faults (fig. 2). On the flanks of the Green Mountain anticlinorium and the smaller

anticline west of Rutland, beds of the east or normal limb dip gently, whereas those of the west or drag limb have steep to overturned dips. This general structural pattern suggests that the deforming movements were that of a couple, with eastern rocks moving westward over western rocks (Brace, 1953, p. 84). Abundant minor structures in the argillaceous and quartzitic Cambro-Ordovician rocks are generally consistent with this major movement pattern. Thus the rotational features (folds and rotated grains) agree with the general symmetry of larger scale flexural folding, and the linear features agree either with the regional axis of folding or with the presumed direction of overall tectonic transport.

The conglomerates have undergone a metamorphism which has produced biotite-chlorite phyllite in enclosing argillaceous rocks and epidote-albite-carbonate greenstone in volcanic rocks.

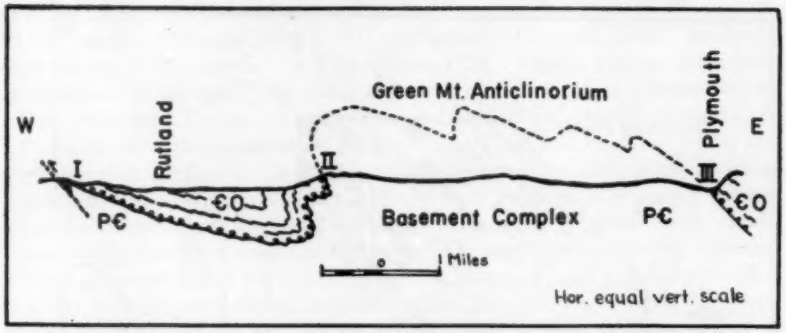


Fig. 2. Geologic cross-section, at the latitude of Rutland, Vt., showing the structural environment of the conglomerates. Pebbles 1, 2, 3, 4, 6 and 7 are from III, pebbles 5 and 8 are from II and pebble 9 is from I. After Brace (1953).

MEGASCOPIC FEATURES OF THE STRETCHED CONGLOMERATE

Description.—The conglomerates are lenticular units attaining thicknesses up to several tens of feet. They are interbedded with quartz-rich grits and graywacke and thin beds of dolomite and sericite-chlorite schist. Rare graded bedding and scour-and-fill structure interrupt the monotony of the thicker beds. The pebbles have all undergone extreme deformation and have the present form of paddles or cigars, 3 to 30 inches long with length 5 to 20 times the thickness. The dark gray-green graywacke matrix ranges from 20 to 80 percent of the volume of the rocks, and is composed of fine-grained quartz, microcline, albite, white mica, chlorite, biotite, epidote and accessories. Foliation in the conglomerates is poor as mica is seldom more than 10 percent by volume.

Some idea of the initial condition of the pebbles[1] is given by Osberg in his description of relatively undeformed pebbles in the trough of a syncline in the East Middlebury area, Vermont (1952, p. 90). These have a shape

[1] The term "pebble" here denotes all distinctly clastic grains larger than a quarter of an inch, a usage common elsewhere in the description of deformed conglomerates. A diametric classification might be preferable but, as the materials were all deformed, mean diameter is not always obvious.

index which is similar to that of modern beach and glacial gravels (Pettijohn, 1949, table 49, p. 201).

The present pebble shapes are approximately ellipsoidal (fig. 3), and their geometry can be related to a framework based on the minor structures in enclosing argillaceous beds. This framework has both planar and linear components. Compositional banding and foliation are *planar* features and are consistently parallel to bedding even in the noses of folds. Slip cleavage is a second *planar* feature developed in micaceous rocks (note side of block, fig. 3) as a result of slip *parallel* to foliation (Brace, 1953, p. 84-91). Slip cleavage is always inclined to bedding and foliation. *Lineation* in foliation surfaces includes crinkles, folds and intersections of foliation with fracture and slip cleavage. These lineations parallel the direction of regional fold axes; this direction is termed the *b* strain axis. A second *lineation* found in foliation

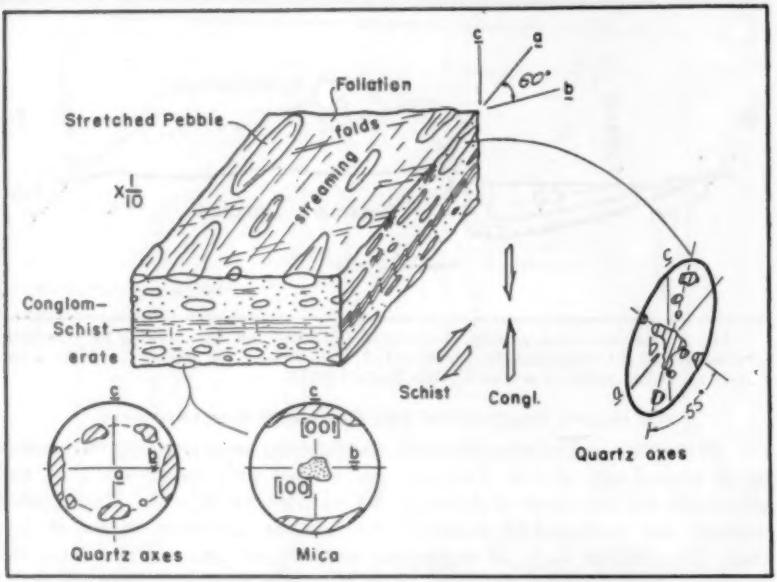


Fig. 3. Megascopic and microscopic features of deformed conglomerate. Fabric diagrams give the typical quartz and mica grain orientation from conglomerates of the Rutland area. Arrows (center) show inferred movement undergone by deformed schist and conglomerate.

surfaces includes quartz pod and pebble elongation, streaming, crenulation and rare folds which are aligned in a direction subnormal to regional fold axes; this lineation defines strain axis *a*, the direction of tectonic transport.

The long and intermediate axes of the stretched pebbles lie in the foliation surface (fig. 3), with the long axis subnormal to fold axes and parallel to streaming. The ratio of major and minor axes ranges from 5 to 20. Many pebbles are not strictly ellipsoids but have lenticular cross-sections, pointed ends or ends of different curvature. In one or two localities tight folding has

produced canoe- and S-shaped forms from the stretched pebbles. The grit matrix of the conglomerates is often strongly crinkled with development of slip cleavage; this axial plane cleavage nowhere parallels the elongation direction of the pebbles but is truncated against the pebble surfaces.

It is of some interest to compare these megascopic features of the stretched pebbles with data reported elsewhere. Pebbles (Oftedahl, 1948, p. 477; Thompson, 1950, p. 104) and fossils (Heim, 1878, p. 61) are commonly flattened or elongate in bedding or in foliation parallel to bedding. Yet this is not universal, for pebbles (Fairbairn, 1936) and ooids (Cloos, 1946, p. 35) may have major and intermediate axes in an axial plane foliation. The orientation of the long axis is even more variable than the planar relationships. In contrast to pebbles from the Rutland area Strand (1944, p. 17), Fairbairn (1936), and Kvale (1947, p. 28) report pebble elongation parallel to fold axis, or fabric *b*. Elongation in *a* is described in Vermont by Thompson (1950) and by Osberg (1952, p. 90), in Norway (Oftedahl, 1948), in Maryland (Cloos, 1946), and in the Alps (Heim, 1878). Clearly then conglomerate deformation of this study is easily matched elsewhere as to orientation of pebble ellipsoids. The explanation of this pattern of deformation remains.

Tentative explanation of pebble shape based on field relations.—The observation that the pebbles lie with long and intermediate axes in the bedding plane may have considerable significance. In a medium undergoing change of shape by simultaneous slip on parallel surfaces, a spherical body should theoretically change to an ellipsoid with long axis inclined to the slip surfaces. This is, of course, the concept of the strain ellipsoid, and it will be recalled that the AB^2 ellipsoid planes should parallel axial planes of folds (Billings, 1942, figs. 183, 184, 189). Now in the stretched pebbles an approximation to ellipsoidal shape is produced from forms which were originally more nearly spherical. The AB^2 plane of the pebbles is invariably inclined to axial planes and is parallel to foliation which has elsewhere served as a surface of shearing movement. Elongation of pebbles has taken place *parallel* to the slip surfaces, the movement on which accounts for nearly all of the other structural features of the area. This elongation parallel to slip surfaces in the surrounding rock might be explained by a combination of slip and rotation, or by the assumption of a deforming mechanism for the pebbles different from that of the enclosing incompetent rock. In explaining a Norwegian occurrence similar to that found in the Rutland area, Oftedahl (148, p. 484) believes that ellipsoidal form is due to homogeneous slip (in the manner of the strain ellipsoid analogy) followed by rotation of the ellipsoidal pebble within the rock matrix into parallelism with the slip surfaces. Now rotation of this sort should leave its mark. The pebbles of this study lie in a foliate matrix with weak banding of granular and micaceous minerals, and this banding is considered primary because it is parallel to distinct quartzose and limy beds. The banding is sufficiently distinct so that its course can be traced through delicate folds and slip cleavage. The pebbles do not appear to have

[2] Axes A, B, and C of either strain ellipsoid or stretched pebbles refer to maximum, intermediate and minimum dimensions respectively. They should not be confused with *strain axes a, b, c* which are based on the geometry of foliation, folds, etc.

locally disturbed these delicate surfaces in the process of becoming elongate parallel with them, although rotation would have been of the order of 30 degrees. Therefore, Oftedahl's hypothesis of slip and subsequent rotation is not considered applicable. The pebbles must have been elongated by some mechanism other than homogeneous slip on parallel surfaces. It is suggested

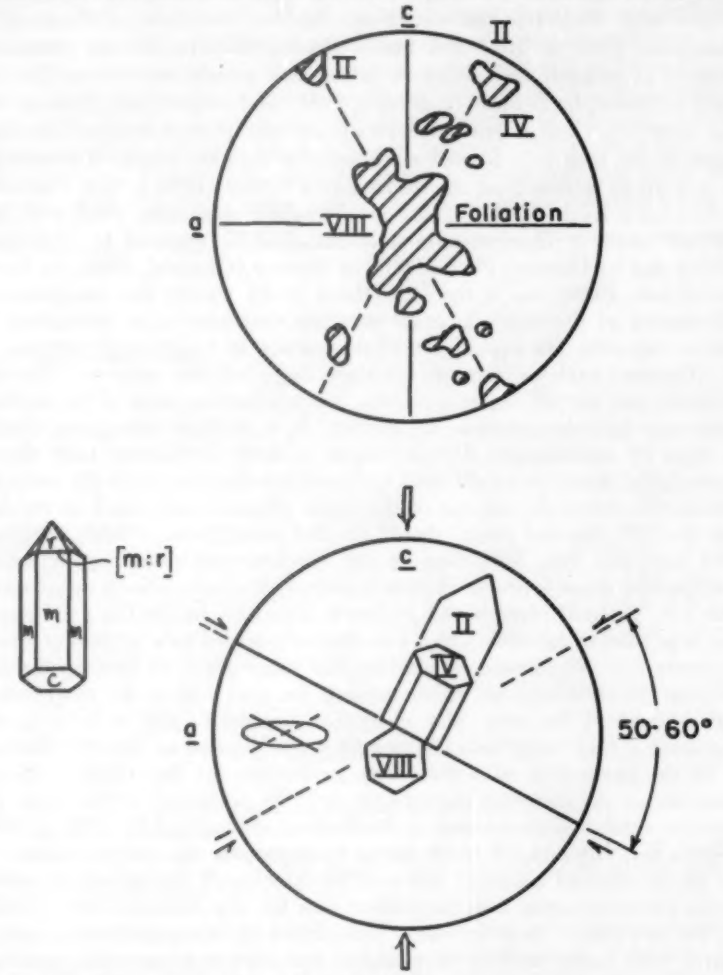


Fig. 4. Quartz orientation in deformed pebbles.

(above) Synoptic diagram of quartz axis orientation in nine stretched pebbles rotated $\perp$ *a*. Roman numerals refer to the location of maxima in coordinate system *a*, *b'*, *c*.

(below) Location of quartz grains in hypothetical slip planes in a pebble, giving the maxima observed in diagram above. Quartz is assumed to undergo translation gliding on c (maxima II), on r (maxima IV) nd on m (maxima VIII), in a direction [m:r]. These faces are sketched on a quartz crystal (left).

that flattening normal to foliation has been important. It remains to examine the microscopic fabric of the deformed pebbles to test this hypothesis.

QUARTZ FABRIC

Deformed pebbles.—The fabric of nine deformed quartzite pebbles was examined; the results are summarized in table 1 and shown in lower hemisphere projection in figure 5. Although it is rather difficult to establish trends among the various fabric elements, they are described below and an attempt made to list those features common to the fabric of most of the pebbles.

In the absence of more complete field data, it was not possible to investigate the relationship between geometrical ratios and the degree of quartz orientation as shown by percent maxima of quartz axes. In the data available, however, it is rather surprising to note that some of the highest percent maxima occur in pebbles with axial ratios ranging from 6 to 1 to 10 to 1, whereas in pebbles with ratios nearer 20 to 1 the degree of quartz orientation is much less marked.

Typically there is a great range of grain size in the quartz of a single pebble (all the pebbles are fragments of quartzite). Large undulose grains are commonly embedded in a matrix of clear, equant, finer-grained quartz. There is no particular agreement between the degree of orientation, as expressed by percent maxima of quartz axes, and either absolute grain size, or the range of grain size. Thus the highest and lowest maxima (8 and 5 respectively, fig. 5) occur in pebbles of about the same texture.

Grain dimensional orientation could be fully evaluated in those few pebbles for which more than one thin section is available. In these, quartz is distinctly elongate parallel to the long axis of the pebble. For the majority of the pebbles only a section normal to A was cut, and in these dimensional orientation is surprisingly weak. Grains are slightly longer parallel to B and simulate the shape of the pebble. Again there appears to be no correlation between the degree of dimensional orientation and the degree of lattice orientation of quartz, although admittedly the data are incomplete.

The presumed effects of strain include undulose extinction, fracturing, and sutured grain boundaries. These effects appear typically in the larger grains within the pebbles and are absent in the equant, smaller, interstitial grains. Lamellae appear in a few of the pebbles and are normally highly oriented in planes making a large angle with the foliation surface or AB plane of the pebble, but too few occur to permit an exhaustive study. Quartz lattice orientation may be slightly better in the visibly most strained rocks but this is at best a weak trend. It is again rather notable that in two of the pebbles (8 and 9, fig. 5) with nearly the largest quartz axis maxima, few visible strain effects are preserved in the fabric.

As might be anticipated, white mica shows strong dimensional orientation within and at the surfaces of the pebbles. This is expressed in concentration of poles to cleavage surfaces [001], which universally lie subnormal to *a* or the long axis of the pebble. In sections cut at right angles to the long dimension of the pebbles the poles to mica grains form incomplete girdles parallel

to the surface of the section. These are $b'c$ girdles[3] which in all likelihood arise from the wrapping of the mica grains about the cigar-shaped pebble forms. The scatter of poles within these $b'c$ girdles ranges from 20 to nearly 360 degrees. If a single small pebble is included within a thin section, the poles to its enclosing mica show a complete girdle; elsewhere, depending on the size and position of the thin section relative to the pebble, they show incomplete or irregular girdles. There was no tendency for mica poles to lie in an ac girdle parallel to the pebble length, but there might have been if sampling had included the many tiny folds which cross the pebble length subnormal to its long axis (fig. 3).

Quartz in the deformed pebbles has the strongest preferred orientation of any of the quartz- or carbonate-bearing rocks examined from central Vermont. Quartz axes are grouped into maxima in concentrations which range from 5 to 16 percent (table 1). One or more of these maxima typically occur in incomplete girdles which intersect in the b' axis of the fabric; these girdles are termed {h0l} girdles with reference to the rectangular coordinate system, a, b' and c. Seven of the nine pebbles have quartz axes which define such girdles; in these seven the average angle between the planes of the girdles is 125 ± 15 degrees, the girdles lying nearly symmetrical with respect to the ab' plane or foliation.

For convenience the locations of maxima within the rectangular system a, b', c are given a Roman numeral designation (see figure 5, diagram 10 for explanation). Six of the pebbles have strong quartz axis maxima which fall near VIII or subparallel to the b' axis of the fabric. Seven of the pebbles have maxima lying in the girdles which fall within the area given the designation IV. Finally, several maxima fall near positions III and II, the latter being the intersection of the girdle with the ac plane. Clearly there is no great consistency of the quartz preferred orientation throughout the nine samples. Thus maxima at VIII may be isolated or they may be associated with maxima at III and IV, and it is rather difficult to detect a symmetry within the orientation patterns which is common to all the pebbles. The only conspicuous features are (1) a spacing of the maxima at about 90 degrees within the girdles or within a common plane, or (2) a symmetrical position of a pair of maxima in a common plane through the a axis. The first is illustrated by pebbles 5, 6, 7, and possibly 3, and the second by 1, 2, 4, and 9.

Comparison of matrix and pebble quartz orientation.—The quartz orientation in the conglomerate matrix of three of the pebbles numbers 4, 5, and 9, was also studied. The fabrics of these are shown in figure 5 along with their respective pebble fabrics. In all cases matrix grain orientation was measured within a few inches of the pebble surface.

There are obvious differences in the fabric of pebbles and matrix. Visible strain phenomena are usually absent in the quartz of the matrix. Furthermore, quartz axis orientation is in most cases markedly different.

[3] Inasmuch as fold axes and the long dimensions of pebbles (axes b and a respectively) nearly always make an angle of 50 to 80 degrees in the foliation surface, an axis b' is used here to permit the convenience of a rectangular coordinate system. Axis b' is perpendicular to a and lies in the foliation surface.

TABLE 1

Pebble	Grain Size	Dimensional Orientation	Strain Phenomena	Lamellae	C_V Quartz Grains	White Mica
1.	0.2 mm 0.5-0.05 mm	none	30% of grains show und. extinction	none	incomplete {h0′l} gdl. 8% max. at II	*b′c* gdl. of [001]; [100] normal to *a*
2.	0.4 mm 0.1-0.8 mm	strong: B>C*	all grains show und. ext.	rare	incomplete {h0′l} gdl. 9% max. at VIII and IV	nearly complete *b′c* gdl.
3.	0.3 mm 0.4-0.05 mm	faint	50% of grains show und. ext.	none	{h0′l} gld.; 5% max. at III and IV	60° spread of [001] in *b′c* gdl.
4.	0.2 mm	strong B>C	pronounced in large grains only	none	incomplete {h0′l} 9% max. at VIII	100° spread of [001] in *b′c* gdl.
5.	1.0-0.04 mm	weak: B>C	"	"	incomplete {h0′l} gdl. 5% max. at III	150° spread in *b′c* gdl.
6.	0.06-0.08 mm	"	none	"	excellent {h0′l} gdl. 7% max. at II and VIII	40° spread of [001] in *b′c* gdl.
7.	10.0-0.05 mm	?	strong und. ext. in all grains	conspic. in large grains; are in 2 planes {0kl} at 90°	two 15% max. in IV	?
8.	0.03-1.0 mm	none	none	none	Single 16% max. in VIII	30° spread of [001] in *b′c* gdl.
9.	0.02-0.8 mm	none	20% show und. ext.	none	two sym. 12% max. in IV	?

* B>C means grains elongate in *b* with respect to *c*.

Note: *b′* is imaginary axis *normal* to *a*.

{h0′l} and {0kl} refer to coordinate system: *a*, *b′* and *c*.

Key to Roman numerals designation is given in figure 5, diagram 10.

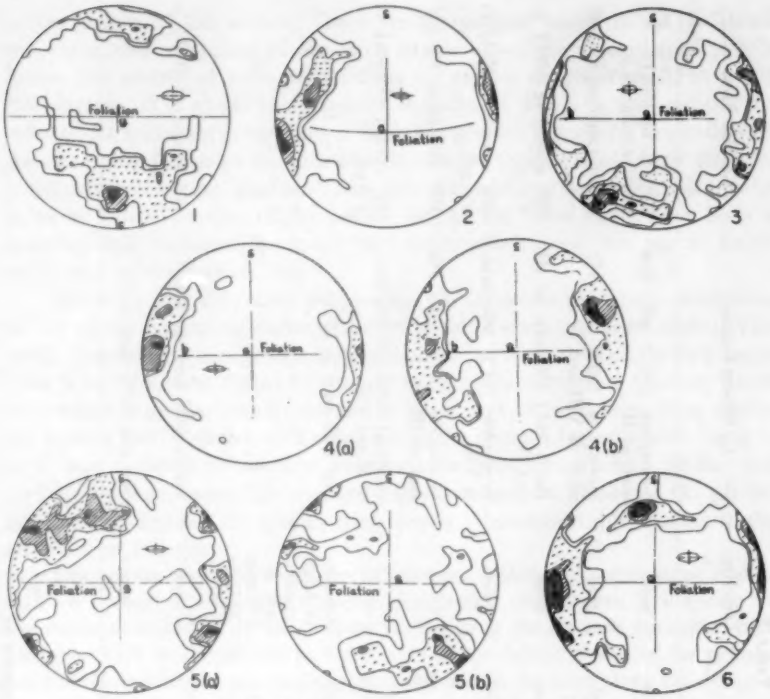
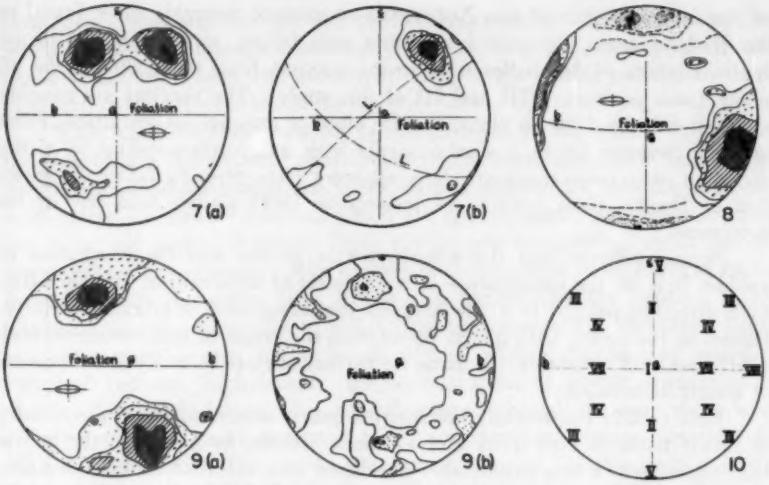


Fig. 5. Fabric diagrams of pebble quartz.

1. Quartz axes from quartzite pebble, Plymouth. Contours: 1-2-4-6-8%.
2. Quartz axes from quartzite pebble, Plymouth. Contours: 2-4-6-8-10%.
3. Quartz axes from quartzite pebble, Plymouth. Contours: 1-2-3-4-5%.
4(a). Quartz axes from quartzite pebble, Plymouth. Contours: 2-4-7-9%.
4(b). Quartz axes from conglomerate matrix of same pebble. Contours: 1-2-4-6-8%.
5(a). Quartz axes from quartzite pebble, Mendon. Contours: 1-2-3-4-5%.
5(b). Quartz axes from conglomerate matrix of same pebble. Contours: 1-2-3-4-5%.
6. Quartz axes from quartzite pebble, Ludlow. Contours: 1-2-4-5-7%.

1. The percent concentration in maxima is much less in the matrix than in the pebble although the location of the maxima in matrix and pebble is approximately the same; in sample 5, however, one marginal maximum (III) of the pebble is entirely absent in the matrix pattern.
2. There is a noticeable scattering within the matrix, a tendency to diffuse the sharp patterns of the pebbles. Quartz axes are less clearly arranged in {h0$\bar{l}$l} girdles in the matrix patterns than in the pebble patterns and some of these axes seem to fall on rather vague *b'c* girdles.

Previous work.—The quartz fabric found here in deformed pebbles has features which are apparently not well known. Elsewhere maximum VIII appears in quartzite associated with a girdle parallel to *ab* (Hietanen, 1938,



7(a). Quartz axes from quartzite pebble, Sherburne. Contours: 2-4-6-10-15%.

7(b). Poles to lamellae from those grains of pebble quartz in which lamellae are visible. The approximate location of lamellae (L) are shown.

8. Quartz axes from quartzite pebble, Bear Mountain, Wallingford. Contours: 1-2-4-10-16%. Poles to mica (0001) cleavage are shown as dotted lines. Contours: 1-5-10-15-25%.

9(a). Quartz axes from quartzite pebble, Pine Hill Proctor. Contours: 2-4-6-8-12%.

9(b). Quartz axes from conglomerate matrix of same pebble. Contours: 1-2-4%.

10. Roman numeral designation of maxima with reference to fabric axes *a*, *b*, and *c*. After Fairbairn (1949, p. 11).

Note: These diagrams represent *lower* hemisphere projections of fabric elements, in most cases oriented so that pebble long axis is perpendicular to the diagram. *a*, *b*, and *c* are strain axes. The universal stage procedure and method of preparation of diagrams is described by Fairbairn (1949, p. 241-290). 250 grains were measured without selection in each section and contouring done with respect to concentrations of axes in 1% areas.

for example). Maximum III is apparently characteristic of granulites from many localities; it occurs with symmetrical {0kl} girdles (Sahama, 1936, for example). Maximum IV occurs rather commonly, usually in girdles subparallel to the *ac* plane (Fairbairn, 1939). Girdles subparallel to the *bc* plane have been reported by Schmidt (Mugl gneiss; 1925), Kvale (1945), Strand (1944) and others, and {h0l} girdles are reported by Balk from quartzite at thrust faults (1952). Thus most of the observed fabric elements of the pebbles have been reported but rarely in the patterns observed here or in rocks at all similar to those here studied.

Strand's investigation (1944) of the Bygdin conglomerate of southern Norway is the only detailed work on the structural petrology of deformed conglomerate known to the writer. The pebbles and their environment in his area differ somewhat from those described here in that (1) deformation has been much more intense (axial ratios range up to 80:1.5:1), (2) faulting has been important nearby and (3) folds are commonly arranged parallel to *a*, the direction of maximum pebble elongation. Nevertheless many features

of the fabric pattern of this Norwegian occurrence resemble those found in the Rutland area. The principal quartz axis fabric, away from the thrust faults, consists of *bc* girdles with strong maxima from 0° to 30° to the *ab* plane (near positions VIII and III of this study). The maxima are roughly symmetrical about the *ab* plane, but are often of unequal concentration. Poles to mica cleavage lie in a similar girdle with axis corresponding to *a*, the direction of tectonic transport. It is apparent from Strand's fabric diagrams that at Bygdin there tend to be incomplete {h0l} girdles in a few of the specimens.

Strand believes that the *a*-folds, *a*-axis girdles and the elongation of pebbles in *a* are the consequence of a single act of deformation with thrusting in a direction parallel to *a*. He suggests the importance of triaxial strain relations in the rocks, with quartz orientation the result of movements on both {0kl} and {h0l} surfaces. He made no further attempt to analyze the process of quartz orientation.

Balk (1952) has recently investigated quartz orientation from the vicinity of thrust faults in New York and Vermont. Briefly, he finds that the quartz orientation, which was presumably developed as a result of faulting, consists of *bc* girdles of quartz axes. Maxima occur in Balk's study at VIII and IV or at angles up to 30 degrees from the *ab* plane. Balk directs attention to several occurrences of incomplete {h0l} girdles with one set of paired girdles intersecting at an angle of about 60 degrees in the *b* axis. Balk carefully describes features of orientation and texture ascribed to faulting, but offers little explanation of their development.

It is of interest to compare Balk's results with those of Strand. Both found, as did the author in the Rutland area, axes concentrations near the *bc* plane and development of maxima VIII, III and IV. Balk attributes this pattern to faulting, whereas at Bygdin the single pebble near a thrust fault shows an *ac* girdle. There is thus fundamental difference in the interpretation of similar fabric, although admittedly the behavior of conglomerate pebbles and tabular quartzite beds may not be identical.

Faulting is not generally important in the vicinity of conglomerate in the Rutland area; indeed the contribution to the development of present pebble form by shearing on a single set of parallel surfaces is believed to be negligible. Rather than accept *bc* girdles as a criterion of faulting (as implied by Balk's work), the writer prefers to follow Strand and seek an explanation for pebble form and fabric in terms of a flattening mechanism involving shear on pairs of symmetrical planes.

The development of quartz orientation.—Before attempting to explain the quartz fabric of the pebbles as due entirely to deforming movements, the importance of fabric elements preserved from earlier stages in the history of the rocks must be evaluated. The pebbles are quartzite fragments and, judging from quartzite in the Precambrian perhaps comparable to the source of these fragments, the pebble quartz must initially have had a high degree of orientation. In considering the preservation of early fabric, it is assumed that any relic fabric elements would not bear a symmetrical relationship with the coordinate system defined by structural elements obviously produced by de-

formation (folds, slip cleavage and pebble elongation itself). This assumption seems reasonable although conceivably in rare cases pre-existing fabric elements might have been fortuitously oriented in the pebble parallel to those produced in the deformation of the conglomerates.

With few exceptions existing fabric elements in the quartz orientation patterns bear a symmetrical relationship to the coordinates *a*, *b'* and *c*. The very common occurrence of maximum VIII is noteworthy, also the less common paired III and IV maxima which are approximately symmetrical about the *ac* or *ab'* planes. Further, {h0l} girdles are symmetrical about the *ab'* plane. While this symmetry with respect to deformation coordinates is by no means perfect, the writer feels justified in considering the existing fabric elements as due principally to deforming movements. On the other hand, it is also possible that many details of the existing fabric may relate to inherited features. Significantly, the four patterns of quartz orientation of samples 1 through 4 have obvious differences although the four rocks were collected within a small area and have presumably undergone identical deforming movements. The only apparent explanation for this is that initial differences in pebble fabric have somewhat influenced the ultimate deformational fabric. Furthermore, within a single pebble fabric, opposing maxima are often of somewhat different percent concentration; stated differently there often appears to have been a preference for the development of orientation in one of several possible directions. This preference might easily relate to pre-existing fabric, at least in terms of the hypothesis of quartz orientation about to be described.

Any hypothesis of quartz orientation of the pebbles must explain the following features:

1. Differences in matrix and pebble fabric.
2. The occurrence of highly oriented lamellae in {h0'l} and {0kl} surfaces.
3. The common orientation of quartz axes in high concentration in positions VIII, III, IV and II.
4. The scattering of quartz axes in inclined {h0'l} girdles.

Several of these features are summarized in figure 4 which includes the maxima and girdle positions of the nine samples referred to a common foliation surface and *a* axis.

Matrix quartz grains and the smaller grains within pebbles typically have weak dimensional orientation and, compared with the overall fabric of the pebble, a low degree of quartz orientation. Theoretical studies of metals (Harker and Parker, 1945) suggest that the shape of these smaller grains may be the product of recrystallization or annealing. Inasmuch as the degree of quartz lattice orientation is typically weaker in these grains, the recrystallization they represent may have contributed little to the principal lattice orientation in the pebbles. Apparently other processes were dominant.

It was suggested above that flattening normal to bedding involving shear on paired symmetrical planes had played some unknown part in the development of the existing pebble shapes. This was based on field analysis of the

strain relations in the rocks. Examination of the quartz fabric of the pebbles lends additional support to this hypothesis. First, the general symmetry of the quartz diagrams is not monoclinic (about the *ac* symmetry plane), whereas monoclinic symmetry is characteristic of the many structures formed through shearing movements, slip cleavage, rotated porphyroblasts and drag folds. Secondly, rare lamellae are oriented in the pebbles in planes inclined at large angles to the foliation. Fairbairn (1949, p. 126) and Sander (1930, p. 176) suggest that lamellae are surfaces of translation, and if so, flattening parallel to foliation might have been achieved by shearing movements along these lamellae.

A mechanical development of the fabric, rather than growth or recrystallization, appears more likely then with deforming movements producing an elongation of the pebbles by shear on surfaces inclined at a large angle to the ab' plane. In order to develop a working hypothesis, the line [m:r] in quartz is assumed to be a glide line with translation occurring parallel to the prism (m), rhombohedron (r), base (c) and other unknown surfaces in the zone about [m:r]. Fairbairn (1939), and Schmidt (1925) have elsewhere successfully applied this hypothetical translation system to the explanation of quartz fabric. Fairbairn has discussed the crystal structure of quartz in terms of potential gliding systems and believes that the line [m:r] is the most favorable gliding direction; it is apparently impossible, however, to assign preferred glide planes.

It is next assumed that pebble elongation has resulted from shearing on paired inclined surfaces in the approximate position of {h0$\bar{l}$l} and {0kl} lamellae planes. As the principal change in pebble shape has apparently been elongation in *a*, the {h0$\bar{l}$l} surfaces are probably of greater significance. The relative movement along the {h0$\bar{l}$l} surfaces was probably in a direction[4] normal to the intersection of the planes (the b' axis), thus the preferred location of quartz grains to permit translation would be with [m:r] in this direction and the surfaces (r), (m) and (c) parallel to the {h0$\bar{l}$l} planes. Assuming {h0$\bar{l}$l} planes at about 30° to the foliation surface, grains of quartz undergoing translation might assume the positions shown in figure 4. Thus if the prism (m) is utilized as a glide-plane, quartz axes would fall at position VIII; if the rhombohedron (r) is the glide plane, axes would fall near position IV, and if the base (c) is the glide plane, the axes would lie in position II. Referring to the synoptic diagram of quartz axes from the pebbles, figure 4 (a), these maxima VIII, IV and II clearly characterize the pebble fabric. Undoubtedly the original positions of the {h0$\bar{l}$l} planes (of which for clarity one is shown in figure 4) were different from that shown; they were probably inclined to the foliation initially at about 45°. With change in pebble shape, from sphere to ellipsoid, {h0$\bar{l}$l} surfaces may have rotated toward parallelism with the plane of maximum elongation.

[4] It is of interest to note that Kvale (1945, p. 213) advocates a movement on paired inclined surfaces in a direction parallel to their intersection. In general, however, within a single act of deformation it is certainly more likely that the movements which have initiated the formation of inclined pairs of planes would be resolved along the planes to slip perpendicular to their intersection.

It is thus possible to explain a development of maxima VIII, III, IV and II based on a hypothetical gliding system and the assumption of slip surfaces in the pebbles parallel to rare lamellae planes. Several important aspects of the fabric remain unexplained. One is the {h0ℓ} girdles, the other is the unequivalence or lack of symmetry so apparent in many of the individual pebble diagrams.

Unequivalence or lack of symmetry of maxima may be due to the influence of initial (preserved) fabric. Little has been said here of supposed processes by which quartz grains might assume positions preferable for translation gliding. This might involve either rotation of grains into preferred positions, or solution-redeposition phenomena favoring the growth of grains suitably oriented for mechanical deformation. In either case initial fabric might exert influence. Grains fortuitously oriented in one of several positions favorable for mechanical deformation would produce a deformational fabric which was unbalanced. The matter is frankly speculative, however, and discussion must await further study of deformed conglomerates.

The formation of the {h0ℓ} girdles, so clearly a feature of the pebble fabric (see synoptic diagram, figure 4), is also little understood; if, however, an analogy can be made with megascopic minor structures, incomplete girdles of linear features can be explained. Streaming, mineral lineation and the axes of minor folds commonly vary in position within foliation surfaces. They have "maxima" which locally define the axes *a* and *b*, but range somewhat about these "maxima" forming "incomplete girdles" parallel to the principal movement plane, the foliation. The scatter of quartz axes within {h0ℓ} slip surfaces might be explained in similar fashion. We are dealing necessarily with inhomogeneous media. Neither vertically nor within bedding surfaces are the layered units of the area uniform. The principal differential movements should be deflected somewhat within slip surfaces to accommodate local differences in physical properties of the rocks. On a smaller scale quartzite pebbles are perhaps less irregular than interbedded phyllite and sandstone, but nevertheless are texturally inhomogeneous. Within {h0ℓ} surfaces it is then not unreasonable to find a scattering of linear elements about a maxima; such a scattering is merely a reflection of the inhomogeneity of the pebbles.

CONCLUSIONS

1. No correlation is possible between degree of pebble elongation and the development of microscopic features such as lamellae, undulose extinction and fracturing. Individual grains are, however, weakly elongate parallel to the long axes of pebbles.

2. Recrystallization and growth may have contributed little to the lattice orientation of the pebble quartz. Recrystallization has undoubtedly been widespread, however, and served to erase many of the visible strain effects and may have weakened lattice orientation.

3. From two independent lines of investigation flattening normal to bedding involving shear on paired symmetrical planes appears to have been

the dominant process which formed the stretched pebbles. The arrangement of the pebbles elongate *parallel* to slip surfaces in the enclosing rock precludes elongation by slip on a single set of surfaces. Quartz grains and ultimately the pebbles have apparently lengthened by slip on pairs of surfaces inclined to bedding. Translation gliding in the quartz grains is suggested in view of the position of lamellae. A hypothetical glide system can be successfully applied in which [m:r] is the glide direction in planes (m), (r), (c) and other unknown surfaces in the zone about [m:r].

This is at best a preliminary study and additional data on pebble geometry and fabric should be brought to bear on the problem. Furthermore the conclusions reached here might well be evaluated for pebbles deformed in somewhat different kinematic environment, giving for example disc- or pencil-shaped forms.

Deformation leading to stretching (or conversely flattening) of known shapes offers an excellent opportunity for correlation of megascopic and microscopic fabric. Deformed fossils and ooids should be investigated in the light of calcite deformation and the data provided by laboratory studies of marbles. For conglomerates the degree of elongation (in a sense, the "plasticity") and degree of mineral orientation might be correlated with the texture and size and, in a general way, with the composition of the fragments.

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RELICT DIKES AND RELICT PSEUDODIKES*

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ABSTRACT. Relics of the host rock are unique features of granitized masses, whereas xenoliths and cognate inclusions are to be expected in granitic bodies of magmatic origin. Elongated, narrow, tabular relics simulate post-granitic dikes, but such features as transecting veinlets of the surrounding granitic rock or irregular embayments of the granitic rock, clearly indicate their pre-granitic age. If these narrow bodies were actually formed as dikes in the host rock prior to granitization, the term relict dike is applicable, but if they are merely long, thin, slab-like relics of the host rock, the term relict pseudodike is more suitable.

Relict dikes and relict pseudodikes might be interpreted as remarkable xenoliths were it not for the convincing evidence that their long unbroken horizontal and vertical extent is in accord with static conditions prevailing during granitization and is not what would be expected if they had been engulfed in a mobile magma. These relict bodies are probably best explained as earlier dikes or bands in the host rock which have acted as resistors to the later widespread granitization and hence have remained as relics in the granitized rock.

INTRODUCTION

Post-granitic dikes are of common occurrence in many granitic areas of either magmatic or metasomatic origin. These dikes may be the result of different mechanisms of formation ranging from those of simple igneous injection to those formed by the complex processes of metasomatism. Less common, but of diagnostic importance concerning the mode of origin of the plutonic mass are dikes or dike-like bodies which display features clearly indicating that they are of pre-granitic age. For purposes of comparison, a brief description of post-granitic dikes is presented before the more detailed discussion of pre-granitic dikes or dike-like bodies.

Orthomagmatic dikes.—Intrusive dikes can be recognized by such well-known features as offsets of units in the wall rock, caused by dilation, where transected at an acute angle, chilled borders, or a change to coarser grain size from the borders of the dike to the central part. Later diabase or basalt dikes which were feeders for the extensive Tertiary basaltic flows are most conspicuous in many areas of the Pacific Northwest. In the Cornucopia mining district, situated in the southeastern part of the rugged Wallowa Mountains of northeastern Oregon, these dikes are from 10 to 50 feet in width and have been traced for thousands of feet. Many of them have intersected earlier-formed gold-quartz veins, and where the intersection is at an angle other than a right angle, the veins have been offset because of the dilation of the dike fissure (Goodspeed, 1940). The amount and direction of offset may be determined accurately where the walls of the dike are essentially parallel and where no faulting has taken place.

Replacement dikes.—These dikes in contrast to dilation dikes show no offsets of transected units of the wall rock, have no chilled borders or systematic change in grain size, are usually quite irregular in outline with marked pinching and swelling, and with contacts which are in part gradational. To explain these features it has been suggested that replacement dikes have been formed by the gradual transformation of the wall rock by chemically active

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solutions controlled by parallel fracture planes or permeable layers. At Cornucopia, Oregon, replacement dikes range from less than an inch to a few feet in width and up to 200 feet or more in length (Goodspeed, 1940). Other dikes at this locality show some of the textural features of replacement dikes but, unlike them, exhibit marked flow structures. These have been termed rheomorphic dikes since it appears that the flowage of metasomatized material developed an intrusive dike (Goodspeed, 1952).

Pre-granitic Dike-like Inclusions in Granitized Rock

In addition to the intrusive and replacement dikes at Cornucopia, there are some narrow tabular bodies of fine-grained porphyritic rock which would on casual inspection be considered to be dikes but which display features definitely proving their pre-granitic age. Mention was made of these "dikes" in an early paper dealing with metasomatic processes (Goodspeed, 1939). The paragraph pertinent to these dikes is as follows: "It is usually assumed that porphyry dikes are formed at a slightly later stage in the crystallization of a granitic stock and are injected as dikes through the already consolidated igneous rock. In this region, where apparently there are porphyry dikes cutting granodioritic rocks, it is amazing to find definite evidence that many dikes are older rather than younger than the granodiorite, and thin sections of their contacts show crystalloblastic extension of the granodiorite into the porphyry. Therefore they are older than the surrounding granodiorite and were not injected as dikes into the granodiorite, but probably represent relict bodies which were formed earlier in the original metamorphic rocks and were less susceptible to granitization. Since many porphyries closely approach granodiorite in texture and mineral composition it might be expected that they would not be amenable to granitization. On the other hand, their wall rocks might be more permeable and might more readily react with the activating solutions of granitization."

Relict dikes.—These tabular relict bodies are termed relict dikes because they are relics which have escaped the later widespread granitization and also because their crystalloblastic textures indicate that in all probability they were originally replacement dikes. An earlier igneous dike, as is apparent in some localities, if more resistant to granitization would remain in the granitized rock as a relict body.

Relict pseudodikes.—Narrow tabular relict bodies of country rock are called relict pseudodikes because although dike-like in appearance they were never definite dikes. In general these bodies exhibit the mineralogical and textural features of metamorphic rocks. Like relict dikes their original composition or later mineralogical change, such as partial feldspathization, rendered them more resistant to complete widespread granitization.

Salient features.—Some of the diagnostic features of relict dike-like inclusions are: (1) transecting replacement veinlets of the surrounding granitic rock; (2) irregular cusp-like embayments of the granitic rock into the relics; (3) formation of irregular appendages or thin elongated ancillary inclusions in the granitic rock; (4) shadow-like borders and locally gradational contacts between the "dike" and the granitic rock; (5) crystalloblastic extensions of

the granitic rock into the dikes. At Cornucopia a few relict dikes are edged with a leucocratic (chiefly feldspar and quartz) border in the granitic rock. Thin sections taken across these borders show irregular projections of the "dike" toward the granitic rock and poikiloblastic textures resulting from the incorporation of the finer-grained minerals of the "dike" by the coarse-grained feldspar of the granitic border.

Pre-granitic Dike-like Inclusions in Granitic Rocks of Magmatic Origin

Where the surrounding granitic rock is of magmatic origin, dike-like pre-granitic bodies are not relics but are unusual forms of xenoliths. Such bodies, instead of retaining their form for long distances as in the case of relict dikes and relict pseudodikes, are commonly fragmented by magmatic flow, and the broken pieces of the original tabular bodies become ordinary xenoliths (Grout, 1937). From a genetic standpoint, dike-like xenoliths bear a similar relation to relict dikes or relict pseudodikes that xenoliths bear to skialiths (Goodspeed, 1948). Long, thin, tabular inclusions, either dike-like xenoliths or relict dikes or relict pseudodikes, may exhibit more definitive features than smaller inclusions such as xenoliths and skialiths.

RELICT DIKES AT CORNUCOPIA, OREGON

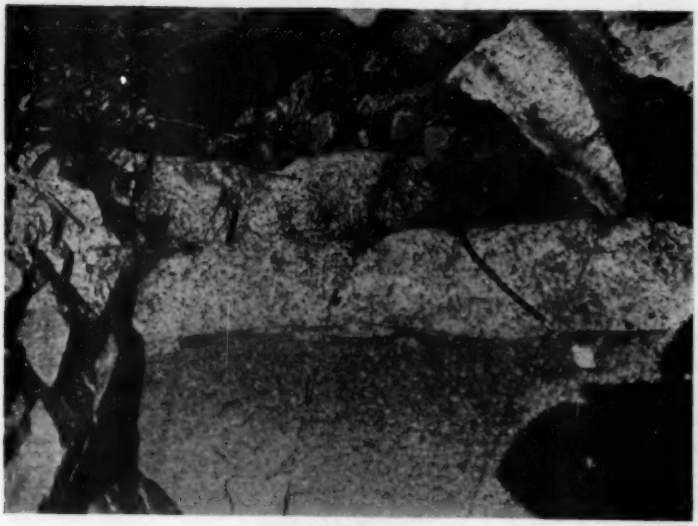
At Cornucopia, the most easily accessible outcrops of relict dikes are on the southern slope of the eastern part of Cornucopia Mountain, an area which lies between two abandoned mining camps, the Union Companion at an elevation of 5800 feet and the Last Chance, at an elevation of 7500 feet. Since much of this area is covered by overburden the best exposures are to be found in road cuts along an old 1½-mile connecting road and in the numerous small gulches. In general, like other pre-Tertiary dikes, the relict dikes average about 3 feet in width and some may be traced for over 200 feet along the strike. They are not as abundant as the replacement and rheomorphic dikes, and were it not for their distinguishing characteristics they could easily be mistaken for replacement and rheomorphic dikes.

Unlike the replacement dikes which in this area occur either in the metamorphic or granitic rocks, the relict dikes are found only in the granitic rocks. Rheomorphic dikes also occur in the granitic rocks, but bear intrusive relations to them, whereas relict dikes exhibit numerous features which show clearly that they pre-date the granitic rocks. The proof that they are older than the surrounding granitic rock and are relics which have resisted granitization is based upon such features as veinlets of granitic rock which transect the relict dikes (pl. 1-A), contacts which indicate a metasomatic invasion by the granitic rock (pl. 1-B), and, as in the case of a few dikes, by a marked feldspathic border zone in the granitic rock against the relict dike (pl. 2-A).

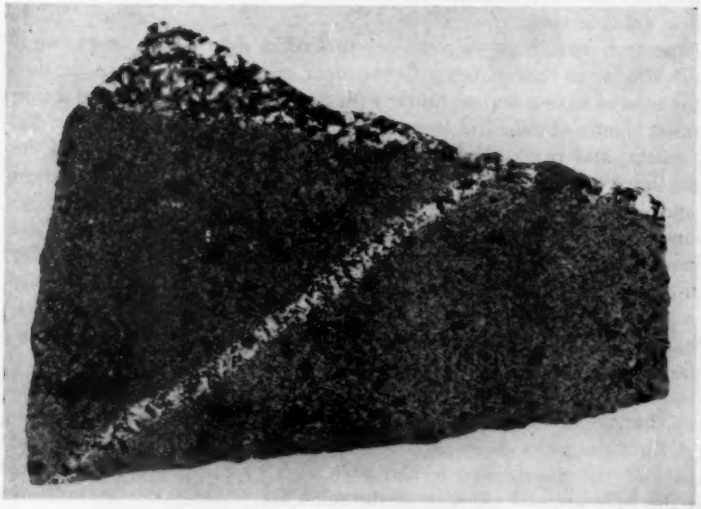
Petrographic Descriptions of Cornucopia Relict Dikes

A good example of a relict dike which is transected by a granitic veinlet is exposed in a cut on the old Last Chance road at an elevation of 6500 feet or about 700 feet above the old Union Camp. This dike is 2 feet wide and has an attitude of N 70° W with a dip of 80° N similar to the schistosity in the hornfels and to the replacement dikes in an area of hornfels a few hundred

PLATE 1



B. Contact of a porphyroblastic relict dike and quartz diorite, Cornucopia, Oregon. Note indentation of the quartz diorite into the relict dike and undisturbed thin appendages included in the granitic rock. Scale is shown by hand lens (1 inch) on relict dike in lower part of photograph.



A. Photograph (crossed polaroid sheets) of a large (3¼ x 4 inches) thin section from a specimen of a relict dike, Cornucopia, Oregon. Note crystalloblastic texture of a transecting granitic veinlet.

feet to the north. It is gray and fine grained with insets of dull feldspar crystals from 0.5 to 3 mm in size, constituting about 10 percent of the rock. Some elongated hornblende crystals and some irregular aggregates of biotite are visible in the hand specimen. A few small irregular cavities (1 to 2 mm in size) are present in this "dike" and also in the granitic rock near the contact. The granitic rock has a medium coarse granoblastic nonlinear fabric and contains easily recognizable plagioclase, quartz and biotite.

In thin section, it can be seen that the groundmass of this relict dike is distinctly xenoblastic, consisting chiefly of small interlocking anhedral grains of plagioclase with some untwinned feldspar which is probably albite and small rounded or sub-rounded grains of quartz. Most of the larger plagioclase crystals, averaging 2 mm in size, are turbid and are subhedral to euhedral. The turbidity is in general caused by the inclusion of very finely divided material which, in some of the plagioclases, can be identified as earlier-formed metamorphic minerals. A few clear feldspars are approximately Ab 63 An 37 in composition. Some of the larger plagioclase crystals are quite anhedral, with irregular outlines which are a characteristic feature of incipient porphyroblasts. Moreover, some of the plagioclases are not single homogeneous crystals but are composed of several individuals of varying orientation. Others have a mottled appearance suggestive of a coalescence of small crystals. These features are common to glomeroblastic aggregates, which represent an intermediate stage of crystalloblastic growth. Conspicuous crystals of hornblende reach a length of 5 mm and hornblendic aggregates over 15 mm are also present. Some of these aggregates are partially replaced by biotite and there are numerous flakes of biotite in the groundmass. The mafics show a rough alignment which is nearly parallel to the contact of the relict dike with the surrounding granitic rock.

The quartz-dioritic rock surrounding this relict dike has a directionless fabric quite similar to that of other Cornucopia granitic occurrences. In thin section it is seen to have a seriate fabric with a maximum crystal size of 2 mm and to consist chiefly of anhedral to subhedral plagioclase (An 40), interstitial anhedral quartz, and ragged flakes or brown and green biotite. Some of the plagioclases have crenulated borders with hairline extensions into adjacent crystals, and a few of them include minute metamorphic minerals with infrequent rounded grains of quartz. Several of the larger plagioclases are not single, simple crystals but are composed of individual crystals which differ from each other in orientation. Those which are single crystals are commonly mottled in such a fashion as to suggest that they were originally aggregates. There are also a few aggregates of biotite quite similar to the biotitic aggregates in the relict dike.

Contacts between the quartz diorite and this relict dike appear to be quite sharp, but under the microscope it can be seen that small flakes of biotite extend from the relict dike into the quartz diorite and also that some of the feldspars of the quartz diorite along the contact have included poikiloblastically the finer-grained minerals of the relict dike. These features indicate that the relict dike was subjected to partial replacement by the same granitization processes by which the quartz diorite was formed and therefore, like

PLATE 2

A. Photograph of a polished surface of specimen showing a contact between a porphyroblastic relict dike and crystalloblastic quartz diorite, Cornucopia, Oregon. Note the roughly scalloped pattern of the narrow leucocratic border of the granitic rock against the relict dike.

B. Photomicrograph of the contact of the leucocratic border and relict dike (pl. 2-A). Note rimmed plagioclase porphyroblasts in dike within 2 mm of leucocratic border and also the scattering of dike minerals into this border. (Crossed nicols.)

the original metamorphic rock, it is older than the quartz diorite. This age relationship is most convincingly demonstrated by the fact that this relict dike is transected by a granitic veinlet which is clearly related to the quartz diorite. This particular veinlet is 2.5 mm wide, has a distinctly crystalloblastic texture, and consists of plagioclase (An 32), quartz, and subordinate biotite. Some of the feldspars of the veinlet span it completely from wall to wall, and some of the biotites cross the veinlet in the form of thin, attenuated aggregates, parallel to the alignment of biotite in the relict dike. There are also numerous ragged irregular projections of the relict dike into the veinlet. All these features suggest that the dominant mechanism of emplacement of this veinlet was metasomatic replacement, controlled by an initial fracture.

Another relict dike exposed in one of the gulches in this area has a porphyritic appearance with numerous plagioclase crystals (1 to 5 mm) set in a very fine-grained slightly schistose groundmass. Some elongated hornblende or biotitic aggregates exhibit a rough alignment parallel to the ill-defined schistosity which in this case is nearly at right angles to the vertical walls of the dike. The surrounding granitic rock has a medium-grained, directionless texture except at the immediate contact with the dike where in a narrow (2 mm) light-colored band the feldspars of the granitic rock have an alignment roughly at right angles to the contact. This band has a crenated appearance similar to the scalloped patterns commonly indicative of replacement aggregates of ore minerals (pl. 2-A). The convex portions of this border band face the dike, and in thin section it is seen that the feldspars of the band include the fine-grained dike minerals poikiloblastically and have locally incorporated or welded themselves to the larger feldspar crystals of the dike. Also in this band small flakes of biotite are interstitial to the aligned feldspars, and in a few instances long thin flaky aggregates of biotite can be seen to extend from the dike into the border band. These features indicate that the granitic rock has partially replaced the dike, while the orientation of the feldspars parallel to the schistosity in the dike suggests a mimetic crystallization to the schistosity (Misch, 1949; Niggli, 1954). Another feature which indicates the permeation of the relict dike by solutions from the granitic mass is the presence of rimmed plagioclase in the dike for about 2 mm along the contact (pl. 2-B). These rims, which have highly crenulated edges, have a lower index of refraction than the original plagioclase, are untwinned, and clear, although filled with inclusions of the finer mineral constituents of the dike. They range from 0.05 to 0.1 mm in width and are probably albitic in composition. Somewhat similar rimmed feldspars were noted by Coombs in his studies of the granitization of the Sauk arkose in the vicinity of Wenatchee, Washington (Coombs, 1950). These features of incipient granitization have been produced synthetically by treating slices of unmetamorphosed Sauk arkose in a bomb for as short a time as 17 days (Smedes and Goodspeed, 1953). The bomb had a capacity of 500 cc and contained 235 cc of solution consisting of 0.16 grams of NH_4CL, 0.66 grams of $NaHCO_3$, and 5.42 grams of $Na_2Si_4O_9$, which gave a concentration of 26.13 grams per liter. Steam pressure in the bomb, although not determined directly, was probably in the order of 4300 pounds per square inch.

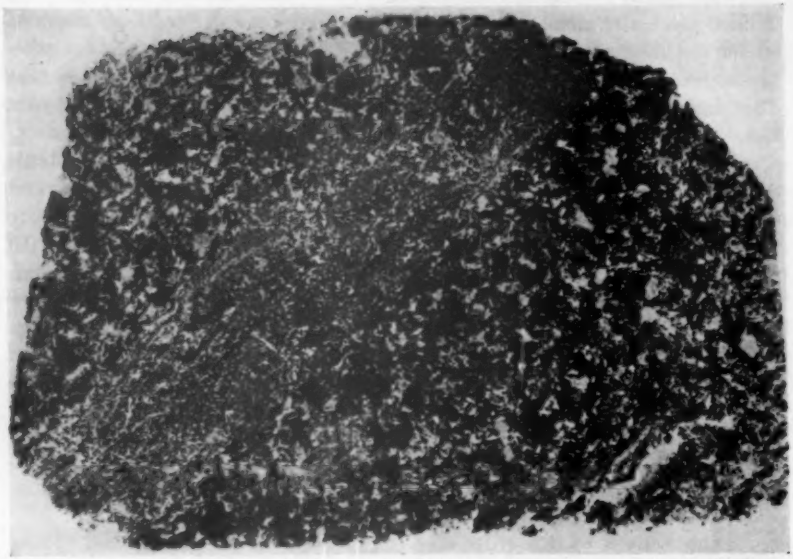
The granitic rock surrounding this relict dike contains some hazy ill-defined patches or small skialiths which may have been derived from the dike. Under the microscope the granitic rock is seen to have a seriate crystalloblastic texture and to consist chiefly of plagioclase (An 45) quartz and biotite. The plagioclases range in size from a fraction of a millimeter to over 4 mm, and the larger subhedral ones commonly have a composite or mottled appearance which suggests glomeroblastic growth. Some show poikiloblastic textures and crenulated borders. The small intergranular feldspars interlock with quartz grains, and in addition to discrete grains of quartz, there are several composite patches of quartz. Irregular flakes of brownish biotite, 0.5 mm in size, occupy an interstitial position to the larger plagioclase crystals of the quartz diorite and locally form rims around these feldspars as if they had been compelled to take this position by the crystalloblastic growth of the feldspars. A few of the mica flakes have been changed from brown to green biotite, suggestive perhaps of an initial stage in the alteration of biotite to chlorite. These petrographic features in addition to the presence of the feldspathic border in the granitic rock next to the relict dike suggest a metasomatic rather than a magmatic interpretation of origin for the surrounding granitic rock. Many other features of the Cornucopia "granites" also indicate granitization as a dominant mechanism of formation.

Thin sections of this relict dike show elongated hornblende and biotite aggregates and numerous larger plagioclase crystals (An 40) ranging from incipient forms which are quite anhedral, with marked crenulated borders, to more euhedral crystals which superficially resemble phenocrysts. Most of the larger crystals, however, contain included material such as easily recognizable grains of earlier-formed metamorphic minerals, chiefly biotite which retains the same orientation as the flakes outside the feldspars. In some feldspars the biotite has been changed to chlorite. Finely divided, unevenly distributed turbid material, consisting chiefly of very minute metamorphic minerals, is also included in the feldspars. Some of the large plagioclase crystals are not simple individuals but complex aggregates. These larger feldspars are set in a fine-grained groundmass consisting of about 40 percent biotite and 60 percent quartz and feldspar with very minor amounts of magnetite and apatite. The petrographic evidence indicates that the larger crystals are porphyroblasts and that this dike was originally formed by metasomatic replacement.

Relict Pseudodikes at Cornucopia

Relict masses of country rock, chiefly schist or hornfels, are common near some of the contacts of the granitic and metamorphic rocks at Cornucopia. Small relics have been termed skialiths so as to differentiate them from xenoliths which are common to igneous rocks (Goodspeed, 1948). It is not unusual to find two large fragment-like relics joined with a relatively thin tabular connecting link which may look like a small dike. For example, the tabular connecting link of a skialith, from a granitic area in the low level adit, is several inches long and 1½ inches wide. As seen in a large thin section it shows much variation with respect both to texture and to mineral

PLATE 3



A. Photograph of a large (3¼ x 4 inches) thin section under crossed polaroid sheets of a thin tabular part of an irregular relict fragment (skialith). Note crystalloblastic texture of surrounding granitic rock and partial feldspathization of the finer-grained relic.

B. Photograph of a large (12 x 14 inches) specimen of a relict pseudodike from the Foster Creek area, central Washington. Note the gradational contacts with the surrounding granitic rock and the irregular patches of feldspathization within the relict pseudodike.

composition in that some parts are composed of finely granular, interlocking, anhedral grains of plagioclase (An 36), whereas other parts display irregular patches and banded aggregates of hornblende with some biotite (pl. 3-A). Some pyrite is strung out with the hornblende and some magnetite is conspicuous in the finer-grained portions. The surrounding granitic rock has a medium coarse-grained seriate crystalloblastic texture and consists of plagioclase (An 37), reaching 4 mm in size, with interstitial quartz and feldspar, as well as some biotite and hornblende. Many of the plagioclases are composite crystals suggesting crystalloblastic growth. This mode of formation is also indicated by some of the mafics, which are bent around the plagioclases.

Along the road, between the upper workings of the Union vein and the bunkers of the Red Jacket Mine, are numerous roughly tabular relics of schist in granitic rock. Some of these schist relics are transected by narrow (1 to 4 inches) tabular bodies, which stand out like ribs on the weathered surface of the schist. These ribs are very fine grained in texture, greenish yellow on the weathered surface but gray on a freshly broken surface, and consist of small irregular feldspars and mafics in about equal proportions. As seen under the microscope, the texture is xenoblastic and slightly schistose, with a rough alignment of some of the mafics and feldspars. The feldspars (0.2 to 0.3 mm) constituting about 52 percent of the rock, are plagioclase (An 32), showing marked sieve structure and crenulated borders. The remainder of the rock consists of green hornblende, diopside, magnetite and a little pyrite. Most of the hornblende is in alignment and the same is true of chain-like strings of diopside; however, some of the hornblende occurs in ragged tuft-like aggregates resembling the pattern of a garbenschiefer. Hornblende also occurs in small joint cracks where it has replaced the adjacent rock to a width of about 2 mm. Some irregular-shaped patches, which are richer in plagioclase than the rest of the rock and leaner in mafics, show a preponderance of green diopside over hornblende.

A few hundred feet north of this locality a similar narrow, tabular, rib-like body of fine-grained, greenish rock occurs surrounded by quartz diorite. It seems probable that the ribs in schist were the result of early metamorphic processes controlled by certain bands in the schist which, by reason of their physical properties or chemical composition, were particularly amenable to recrystallization. It would be expected that these tightly recrystallized bands would be less susceptible to the later processes of granitization and therefore could survive as tabular relics resembling relict dikes. Where there is a strong similarity to an actual dike, the term relict pseudodike is appropriate.

RELICT PSEUDODIKE FROM FOSTER CREEK AREA, CENTRAL WASHINGTON

In the Foster Creek area of east-central Washington are tabular bodies of dark rock which at first sight look like dikes in granite. Their contacts with the surrounding granitic rock are in part fairly sharp and in part quite irregular with embayments of the granite into the dark rock. Their pre-granitic age is shown by the embayments of and transecting veinlets of granitic rock. Their relict nature is plainly evident by the very thin connecting links between embayments, by the presence of thin, undisturbed inclusions, and by the shadowy border zones indicative of incomplete static granitization (pl. 4).

One of these "dikes" which is exposed vertically for about 30 feet in a small quarry dips at a high angle, is about 4½ inches wide, is very fine grained in texture, shows some shadowy borders and contains several elongated, lighter-colored feldspathic patches (pl. 3-B). In a large thin section it can be seen that these patches grade into the darker-colored rock which is a quartz, feldspar, biotite schist, and that they have developed, in general, along the schistosity parallel to the tabular body. They have a crystalloblastic texture and merge into the surrounding granitic rock, which also has a crystalloblastic texture.

It is clear that these tabular bodies are older than the surrounding granitic rock. Therefore, if the granitic mass is of magmatic origin they can be interpreted as peculiar xenoliths or as the downward extremities of roof pendants. If the granitic rock is of metamorphic origin, the simple interpretation is that they are relics of the original country rock which have escaped granitization. The crystalloblastic texture of the granitic rock and the presence of many gradational contacts in this vicinity point toward a metamorphic rather than a magmatic origin for the surrounding granitic rock.

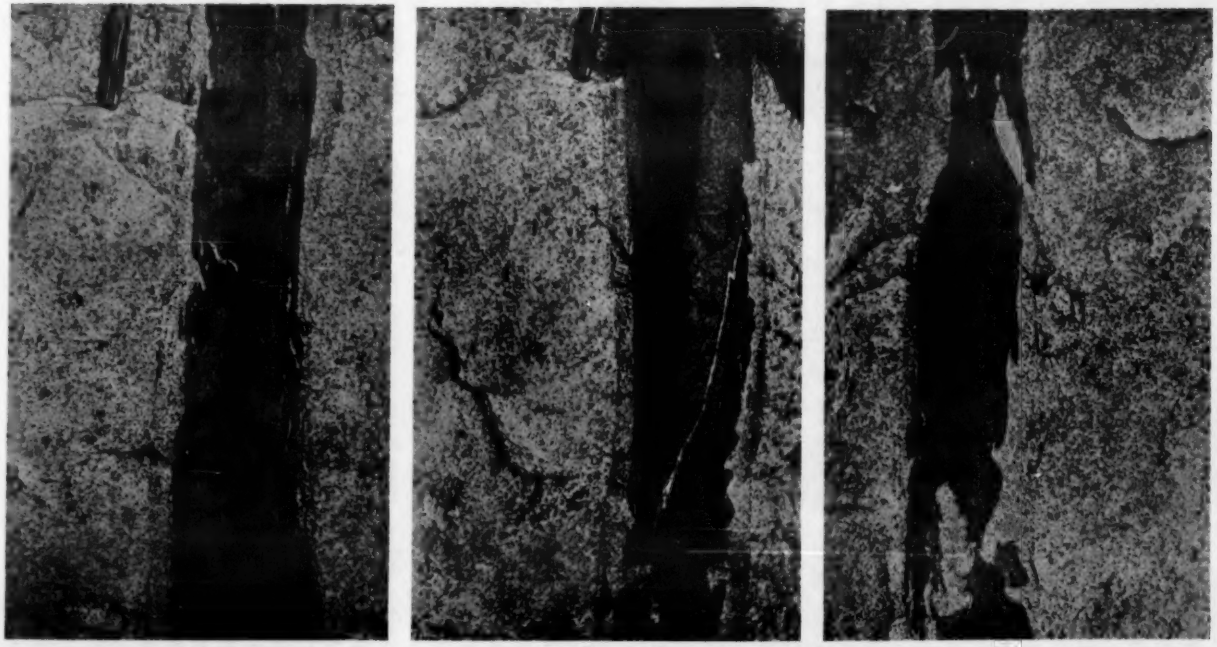
PRE-GRANITIC DIKE-LIKE BODIES IN OTHER LOCALITIES

Dike-like bodies similar to these and also to those at Cornucopia have been described in geological literature, and various terms such as dike-like xenoliths, pre-granodiorite dikes, and pseudodikes, as well as relict dikes, have been used. It is noteworthy that although in general, as might be expected, a magmatic origin for the surrounding granitic rock was given first consideration, yet in some cases the possibility of other processes was mentioned by various authors.

In his report on the geology of Essex County, Massachusetts, Clapp (1921, p. 115-116), stated with regard to dike-like xenoliths: "Many of the xenoliths have roughly parallel sides and are much longer than they are wide and hence resemble dikes. Some of these are so completely broken through and dismembered by the matrix that their included nature is undeniable. The larger ones may be due to an infolding of the roof of the alkaline batholith, but most of them are probably due to the shell-like character of the original fragment spalled from the roof." With regard to the contact action of this batholith, Clapp also states: "Besides the formation of the typical hybrid rocks, another feature of the contact action in Essex County is the infiltration of material from the alkaline magma into relatively cold subalkaline rocks. The nature of this infiltration is similar to that recognized and described by many geologists, in particular by those of the French School."

Callaghan (1935) described pre-granodiorite dikes in granodiorite and carbonate rock in the Paradise Range, Nevada and offered the following statement as an interpretation for some of the older andesite hornfelses and andesite dikes: "In so far as the dikes are attached to the carbonate rock they are roof pendants, but doubtlessly many pieces are wholly unattached. The preservation of original orientation by inclusions is not an extremely uncommon feature in other areas. As the granodiorite follows the earlier dikes in places, it does not seem possible to regard the dikes as septae between separate in-

PLATE 4



A. B. C.

Relict pseudodike, Foster Creek area, central Washington, showing: transecting veinlets (A, B); replacement embayments of granitic rock with thin relict connecting links (B, C); shadowy border bands representing incomplete granitization (A, B); and very thin undisturbed skialiths (A, B, C). Photographs by R. E. Fuller.

trusions that came in, first on one side of the dike and then on the other. If the dikes originally intruded carbonate rocks, then the latter must have been removed by the granodiorite. Some allowance may be made for forcing apart of the walls, yet the disappearance of much carbonate rock must be accounted for. It did not remain in the granodiorite as fragments, nor was it incorporated chemically for the chemical composition of the granodiorite is in no way unusual. It is suggested that magmatic solutions moved in advance of the magma, dissolved the carbonate rocks, and that these chemically modified solutions migrated outward through the country rock or up along the contact. Perhaps this may account for the enormous increment in magnesium in the remaining carbonate rocks in the form of very large deposits of magnesite and brucite and recrystallized dolomite. The calcium is not accounted for but must have been carried still farther away."

In a paper entitled "Observations on pseudo-dikes and foliated dikes," Miller (1945) stated that: "Unusually deceiving are certain long straight narrow xenoliths, which upon casual examination would be called 'dikes'." Although some of the pseudodikes of regular outline described by Miller, especially those surrounded by gabbroid rocks, are no doubt xenoliths, it seems to the writer that others could be best explained as relics. Many of Miller's excellent photographs show irregular networks of pseudodikes joined by relatively thin links. It is difficult to see how such features could remain intact during an orthomagmatic intrusion.

Recently J. E. Armstrong (1950) has noted numerous relict dikes in an area north of Vancouver, British Columbia. In the descriptive notes accompanying the preliminary map, he states: "The Texada group is cut by pre-granite, post-early metamorphic dykes of quartz prophyry, dacite, and granulite. These dykes, most of which strike about 10 degrees west of north or 20 degrees east of north, are found in many places as relict dykes in the Coast intrusions, partly or wholly preserved, and maintaining their original trends, there is little doubt that they are older than the intrusions, as many of them were seen to be cut by the granitic rocks."

He also states: "Included rock, derived from Texada group formations and pre-granitic dykes, may form as much as 50 percent of the plutonic masses, but generally does not exceed 2 to 3 percent, and in places, particularly in the more acidic bodies, may be lacking. These inclusions are commonly round or ovoid but may be elongated. Although these inclusions show all stages of conversions from schist, gneiss, meta-andesite, and pre-granitic dike to plutonic rock, they are dominantly dark grey-green, fine grained rocks of dioritic appearance, containing large porphyroblasts of plagioclase feldspar. These feldspars are similar in all respects to those found in the enveloping plutonic rock."

Pre- and post-plutonic rocks were noted by Harker in his study of the plutonic region of Skye (Harker and Clough, 1904). In a recent paper on the Creag Strollamas area of Skye King (1953) mentions that in one locality basic dikes enclosed in granite show related metasomatism and that in another locality "a short basaltic dyke is entirely surrounded by granite and shows considerable marginal acidification and granite veining." He also makes

the following statements with regard to the mode of origin of the plutonic rocks of this area:

"Most previous geological studies in the British Tertiary region accepted as axiomatic the notion that all igneous bodies have been emplaced by the upward passage of magma from some hypothetical reservoir in depths. Harker, in the Skye Memoir, shows a clear appreciation of the difficulty that is encountered by the application of a 'hypothesis of mechanical magmatic emplacement.' Principally in order to minimize the space problem, he maintained that both the granite and the gabbro 'are generally in the nature of sheets or laccoliths' since 'the only alternative would require us to suppose that the igneous magma has not merely displaced but replaced (by actually incorporating) the rocks into which it has been forced' (1904, p. 83).

"The present account is an attempt at an interpretation of the mode of genesis of the plutonic bodies of the Creag Strollamus and adjacent areas based only on objective evidence. The preservation virtually undisturbed of pre-existing structures of a complicated but easily recognisable character makes it difficult to suppose that the later plutonic rocks were emplaced as magmas from elsewhere by mechanical intrusion. The only alternative conclusion is that the gabbros were formed by the transformation of Tertiary volcanics, while the granites were largely developed by metasomatism of the underlying Torridonian rocks." This means that in harmony with the above interpretation, as given by King for the plutonic rocks, the pre-plutonic dikes are in the nature of relict dikes, which have resisted the widespread metasomatism in this area.

CONCLUSIONS

Because relict dikes and relict pseudodikes can be confused easily with ordinary igneous dikes or with replacement and rheomorphic dikes a summary of some of the diagnostic features is given in table 1. It is to be noted that some of these criteria may apply to more than one kind of dike; hence, in general, several criteria are needed to determine the identity of a particular dike.

The relict dikes at Cornucopia were probably replacement dikes, formed by structurally controlled feldspathization in the metamorphic rocks during an early stage of granitization. At the time of later widespread granitization, these replacement dikes which had previously been transformed from schistose hornfels to a less permeable igneous-appearing rock were not readily susceptible to further metasomatic changes and therefore remained as relics while the metamorphic rocks in which they had been formed were completely granitized. In other localities it is apparent that some relict dikes were early igneous dikes. For these a similar interpretation can be advanced, namely, that they were more resistant to granitization than their enclosing rocks.

In contrast to relict dikes, relict pseudodikes commonly display textural, structural, and compositional features which are usually similar to those of the adjacent regionally metamorphosed rocks. Like relict dikes they are bands which were more resistant to granitization than the bulk of the rock which was transformed from metamorphic rock into granitic rock. Similar to the

TABLE 1

Some of the Features of Dilation Dikes, Replacement Dikes, Relict Dikes, and Relict Pseudodikes

	Dilation: Orthomagmatic and Rheomorphic Dikes	Replacement, Metasomatic Dikes	Relict Dikes	Relict Pseudodikes
1.	Post-wall rock age clearly shown by intrusive relations including the geometrical evidence of dilation, chilled or finer-grained borders in the igneous rock, and varying degrees of optalic metamorphism of wall rocks next to contact of dike.	Post-wall rock age shown by undisturbed included relics of wall rock and crystalloblastic extension into walls. No regularity of textural change, no chilled borders. Lack of geometric evidence of dilation is indicative of a replacement rather than an intrusive mechanism.	Pre-wall rock age shown by transecting veinlets of granitic wall rock and crystalloblastic extension of wall rock into relict dike.	Pre-wall rock age shown by crystalloblastic extension of granitic wall rock into relict pseudodike by cross-cutting granitic veinlets. Metamorphic structures of the relict pseudodike are usually characteristic.
2.	Wall rocks may be of any kind with usually a clear demarcation between the dike and the walls.	Wall rocks may be of any kind with gradational or sharp borders.	Wall rocks are always granitic with sharp or gradational contacts.	Wall rocks are always granitic with gradational or sharp borders.
3.	Uniform composition may extend for great distances.	Composition may display marked changes within a few inches.	Composition will indicate either an earlier magmatic or replacement dike.	Composition depends upon the character of the parent metamorphic rock.
4.	Apophyses are of similar composition to the parent dike although the texture may be finer grained.	Apophyses may differ from the main dike in both texture and mineral composition.	Relict appendages may look like apophyses and may also become detached inclusions in the granitic rock.	No apophyses possible although relict appendages may look like apophyses and may also become detached inclusions in the granitic rock.
5.	Essentially uniform widths may prevail for thousands of feet although irregularities in attitude and width may prevail where dikes have encountered disturbed zones.	Pinching and swelling with extreme variation in width are common. Usually horizontal and vertical extent does not exceed a few hundred feet.	In general uniform widths are characteristic. Irregularities usually caused by later metasomatic action.	Widths may be uniform or irregular depending upon the original character or upon the extent of later metasomatic replacement.

recognition of the genesis of inclusions, namely, whether they are xenoliths or skialiths, the distinction between a dike-like xenolith and a relict dike or a relict pseudodike will add to the criteria as to whether the surrounding granitic rock is of igneous or metasomatic origin.

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SURFACE TENSION AS A FACTOR IN GRADATION

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ABSTRACT. Surface tension and related phenomena of capillarity, adhesion, and cohesion are significant factors in gradation which have not been well recognized. Erosional effects are well shown in gully heads in southeastern United States where soil profiles have a resistant subsoil overlying a friable decayed saprolite of feldspathic crystalline rock. Part of the water which passes as free flow over the rim of a gully head is diverted by surface tension and moves downward or backward along the face of the resistant subsoil until it soaks into the saprolite. This movement is called "backtrickle." The wetted material collapses when saturated and develops a recess called a "cave-head," which eventually undermines the resistant material and causes retreat of the gully head. A similar diversion of the water downward at the lip of an overfall and the accompanying backtrickle develop backsloping boxheads and degrade the sidewalls of canyons in thick sandstones in western United States.

Aggradation occurs when "painting" takes place. Water held by surface tension to vertical or overhanging faces or to steeply sloping surfaces of poorly consolidated material will transport silt and clay downward along a narrow rill channel as a minature mudflow. When the water is absorbed or evaporated the silt and clay are deposited as a small terminal globule called a "muddrip." Hundreds of muddrips and rill channels coalesce and overlap, and when dry harden into a protective crust that prevents sloughing and is resistant to most hard rains. Painting is a feature commonly seen in most parts of the United States. Stains and streaks on vertical and overhanging cliff and canyon walls are results of seepage, surface tension, and precipitation of dissolved minerals.

Features due to degradation and aggradation by surface tension are common. Locally the results may be small, but in the aggregate they are large.

INTRODUCTION

Surface tension is a significant factor in gradation which has not been well recognized. Surface tension and the related molecular forces of adhesion, cohesion, and capillarity affect erosion as well as deposition and resistance to erosion. The individual magnitude of these phenomena is not large, but their wide distribution and the aggregate results are considerable. The effects of surface tension are most commonly found and may be seen nearly everywhere on vertical or overhanging walls or heads of valleys, bluffs, or man-made cuts. They are most pronounced where poorly consolidated beds or deposits alternate with more indurated layers, where beds are silty or clayey, and in canyons and channels through thick poorly cemented sandstones.

Gregory (1917, p. 132-134) noted how the surface tension of thin films of flowing water influenced the erosion of sandstone ledges in boxhead canyons in the Navajo country. Similar boxheads have been studied by the writer in the Caddo Canyons of Oklahoma (Ireland, 1949). Bryan (1925) recognized that surface tension was effective in the formation of "hoodoo" rocks and pedestal rocks in the Southwest. Surface tension is one of the most important factors in gully erosion in the Piedmont region of the southeastern states

A. Gully with cave-head at terrace outlet near Spartanburg, S. C. Cave widens at the rear and is in friable saprolite overlain by resistant clayey subsoil. Note fallen material at the back.

B. Cave-head developed in friable saprolite beneath a quartz vein which separates B and C soil horizons. Note the "painting" of red material adhering to the wall below the vein.

C. Painting process forming a new crust over porous saprolite in South Carolina. Note mudrills and muddrips. (Taken from Ireland et al., 1939).

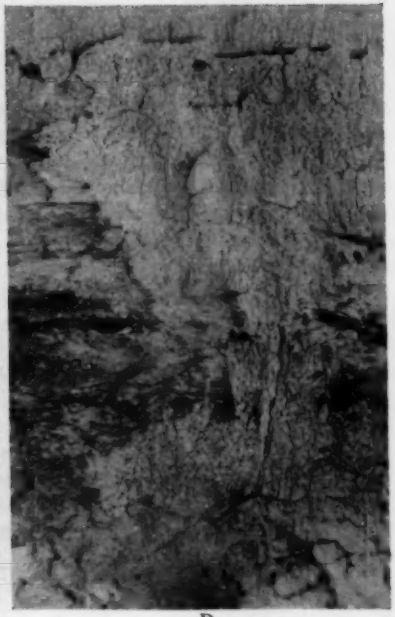
D. Painting on bank of Hope Creek west of Wellborn, Brazos County, Texas.

PLATE 1
Features due to surface tension

A.

B.

C.



D.

because of the greater rainfall and the presence of soil and parent material more susceptible to erosion (Ireland, Sharpe, and Eargle, 1939).

EFFECTS IN THE PIEDMONT

A detailed study of gully erosion in the Piedmont near Spartanburg, South Carolina (Ireland, Sharpe, and Eargle, 1939) led to an understanding and an awareness of the significance of surface tension as an agent of gradation. In this area the Cecil soil series, which occurs throughout the Piedmont, has a subsoil of red resistant sandy clay underlain by a decayed, friable, highly erosible material called *saprolite*, derived from the decomposition of feldspathic igneous and metamorphic rocks (pl. 1-A, B, and C). This soil is probably the most easily eroded in all of the Piedmont. Most of the gullies have been eroded headward along farm-terrace outlets and drainage ditches or along old ruts or ditches of abandoned roads. These channels concentrate water and deliver it to the gully head over a fall. Undermining and collapse of the gully head lengthens the gully. Most of the gullies started before present methods of soil conservation were developed and many of them are more than 80 years old. The sequence in enlargement of one of these near Milledgeville, Georgia, first observed by Lyell in 1842, has been described by the writer (Ireland, 1939).

How surface tension assists erosion can best be illustrated by describing the events in a typical gully in Cecil sandy loam near Spartanburg, S. C. There the runoff from several acres is delivered into a vertical-walled gully with a head having a depth of about 20 feet (pl. 1-A). During a short intense rain, runoff begins to flow over the lip of the gully head (fig. 1). The momentum of the water carries it beyond the lip and it falls to the floor of the gully and scours a plunge basin. However, surface tension diverts a portion of the water at the lip, and this water trickles downward along the face of the resistant subsoil exposed in the gully head. Because the subsoil, or B soil horizon, is clayey and non-absorbent, the diverted water continues downward until it encounters the decayed permeable material of the C soil horizon or saprolite. The saprolite absorbs the water and when saturated falls to the floor of the gully head as "mud." With continued wetting and collapse of the material of the saprolite a recess or cave is developed, overhung by the resistant material of the B soil horizon. A gully head of this type is called a *cave-head*. Surface tension continues to operate, and the cave grows larger as water trickles downward and backward along the roof of the cave. This movement of water is called *backtrickle*. In many places water may collect in such large drops on the roof of the cave that it overcomes surface tension and drips to the floor. Bryan (1925) called this feature *drip curtain* when referring to pedestal rocks, and the name is applicable here. Eventually the overhanging roof is so undermined that it collapses as a mass or several lumps, causing the gully head to retreat a few inches or several feet. A single intense rain may be sufficient to cause a gully head to retreat a foot or more, or several rains may be required. The contact of the B and C soil horizons may be readily differentiated by the point where the roof of the cave-head begins (pl. 1-A, B).

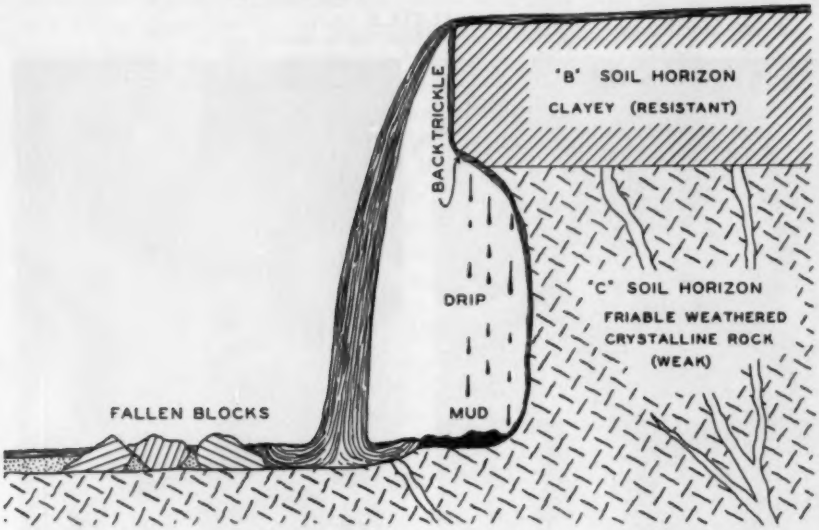


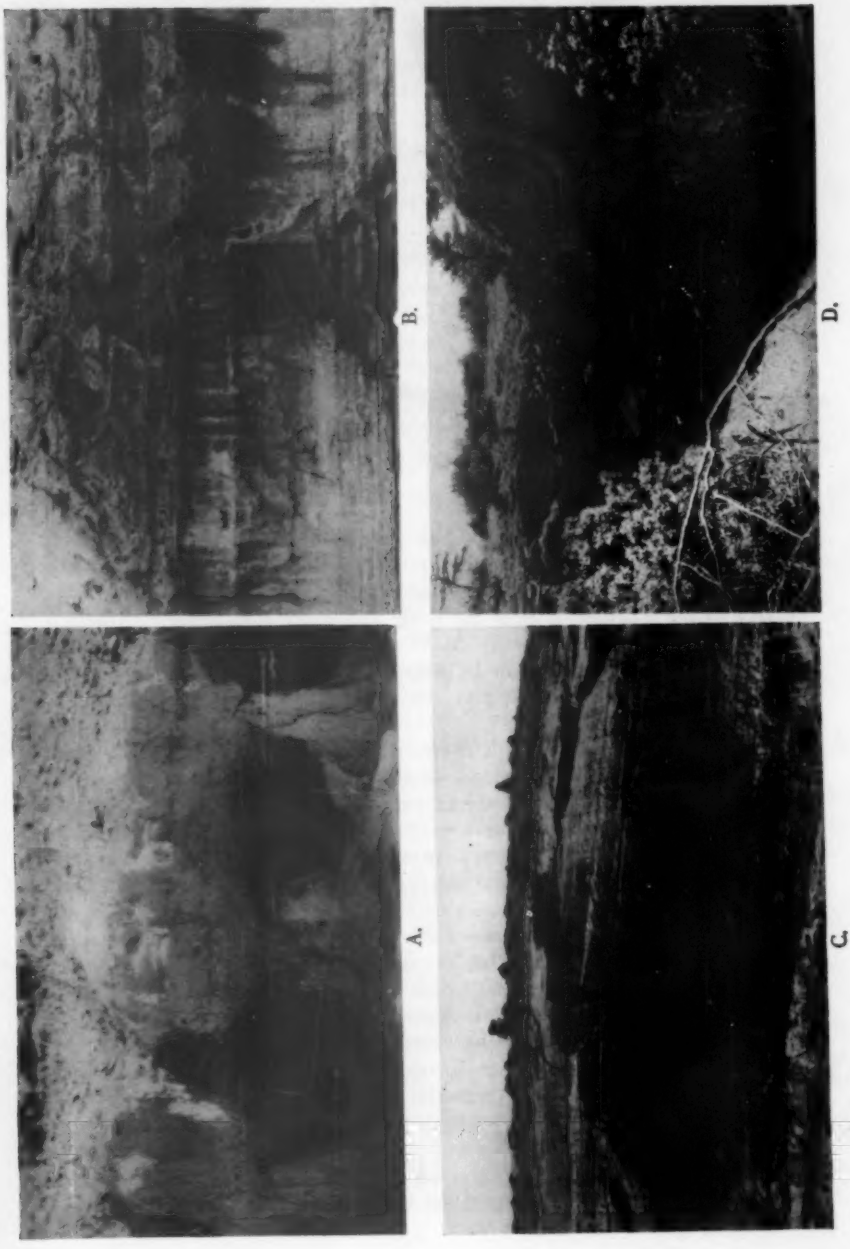
Fig. 1. Diagram showing the operation of surface tension in developing a cave-head.

Surface tension operates, and the process of undermining and caving continues as long as water passes over the lip of the gully head. Even after the overfall of the main runoff has ceased, water may continue to move down the face of the gully head, held there by surface tension, as long as seepage from the preceding rain persists in the main channel. In areas with ample humus and root mats there may be seepage for a considerable time after the rain has ceased. Thus erosion may continue for some time after the main volume of runoff has passed.

A large volume of water with overfall into the gully is not necessary for the action of surface tension. Long-continued rains of low intensity of the warm-front type characteristic of fall and winter in the Piedmont may not supply much runoff, but provide seepage or a minute flow which is held to the gully walls and heads entirely by surface tension, thus allowing backtrickle into the C horizon and the enlargement of recesses. Warm-front rains also tend to saturate slowly the overhanging B horizon of the gully head and sidewalls, leading to their collapse. Gully erosion is likely to be more rapid in late winter, and sometimes in late summer, when abundant and occasionally torrential rainfall of convectional origin removes the collapsed material accumulated during the periods of low intensity rains.

Though surface tension is the chief agent in causing the recession of gully heads, it could not continue to operate if the free flow of water over the lip did not remove the fallen material resulting from backtrickle and drip. The velocity and energy of the water falling into the gully is large, but that energy is dissipated largely in deepening the gully bottom and removing the caved material, rather than in cutting the gully head. Collapsed material may be washed away by the agitation of the water in the plunge basin, or it may

PLATE 2
Painting, stains, and boxhead due to surface tension



accumulate in such quantities that the cave-head or gully sides may be covered to a height above the top of the vulnerable C horizon. Covering of the C horizon stops erosion by preventing backtrickle into the saprolite.

The amount of erosion caused by surface tension may be illustrated by some actual measurements. Datum points were established by setting out stakes at 5-foot intervals along drainage channels above a large number of selected gully heads in South Carolina. Head B of Cox Gully with a drainage basin of only 1.5 acres retreated 20 feet in 23 weeks between October and April. Head A of Littlejohn Gully draining about 6 acres caved 10 feet in 12 weeks from late October to late January. Head A of Foster's Tavern Gully with a drainage basin of 6.5 acres retreated 13 feet in 19 weeks. Much larger amounts of caving occurred, but much of it was due to undercutting by heavy runoff, and such could not be differentiated from that due to surface tension. Records of the rates of cutting of many other gully heads have been published (Ireland, Sharpe, and Eargle, 1939, p. 79-90).

The principles described above may be applied throughout the United States where similar conditions exist in soils or loosely consolidated sediments.

AGGRADATION BY PAINTING

"Painting" is an aggrading effect of surface tension in contrast to the degrading effect of backtrickle. It is widespread throughout the United States. The process was first studied in the Piedmont where it is associated with the development of gullies. Water held by surface tension to the face, roof, and back of the cave-head of a gully moves in numerous small sheets which diverge into miniature mudflows and mudrills. The rills are generally less than half an inch wide and show shallow depressed channels which terminate as globules of mud where the water dripped off leaving its load of silt and clay adhering to the wall (pl. 1-C,D). The globules are called *muddrips*. The globular terminations also develop where the water evaporates or is absorbed. Roofs and walls of gully heads may be covered with an intricate reticulate pattern of coalesced and overlapped mudrills, muddrips, and deposits of silt and clay; this process is called *painting*. The blanket of silt and clay becomes so hard when dry that it resists removal and effectively prevents disaggregation, collapse of material, and the falling of small particles. A dry painted surface generally becomes so tightly bonded and resistant that most rains fail to remove the coating, but may spread the material more evenly and seal the surface more effectively. The coating is destroyed only when undercutting and collapse of the cave-heads or walls occurs. Painting is also developed by sheetflow of water over the rims of the sidewalls of gullies, cuts in stream channels, roadcuts, or other vertical or overhanging banks (pl. 2-A). The

A. Cave-heads and painted surface along abandoned roadcut of U. S. 75 at Bedias Creek bridge, 10 miles south of Madisonville, Texas.

B. Stains and streaks on vertical and overhanging walls near mouth of Pecos River canyon west of Del Rio, Texas.

C. Stains and streaks on Tensleep sandstone in Middle Popo Agie canyon west of Lander, Wyoming.

D. Backsloping boxhead of a vertical-walled canyon in soft Permian sandstone in one of Caddo Canyons southeast of Hinton, Okla.

term "painting" is very appropriate when one observes the great contrast of red, brownish, or other dark material deposited on the face of a light-colored freshly collapsed wall (pl. 1-B, C).

Painting is not confined to vertical or overhanging walls. It is also very common on high-angle slopes in silty and clayey beds. This is especially true in poorly consolidated Tertiary deposits where badland and "cathedral" features develop. Surface tension operates but the effect is less apparent. Muddrips are less common and not as globular on slopes as on vertical walls, but the painting is just as effective. Painting on slopes has been observed widely and it is generally common in beds of all ages where poorly consolidated clay and silt beds occur.

Painting is not likely to occur unless material of clay and silt sizes is present. Clay is necessary for a cohesive bond when the water is absorbed or evaporated. Loess forms vertical walls, but rarely is painted because it lacks clay. Most sandy material fails to cohere tightly enough to form mudrills or muddrips after the water leaves. Painting might develop in very fine sand, but generally it is not bonded sufficiently to resist erosion and is not likely to be seen. Experiments carried out in South Carolina reproduced artificially the formation and development of minature mudflows, mudrills, muddrips, and painting (Ireland, Sharpe, and Eargle, 1939).

Streaks and stains of mineral matter on vertical and overhanging walls and bluffs are common features resulting from the action of surface tension which may be seen nearly everywhere. Nearly every geologist has seen them, but probably few have been aware of their close connection to surface tension. They are a form of painting due to chemical precipitation rather than particle deposition. No aggradation or degradation is obvious, but very slow chemical weathering may occur in some places. Lichen and minute plants may appear to cause the discoloration, but their presence in most cases is a result of moisture made available by the action of surface tension. Bluffs of the canyons of the Pecos and Rio Grande rivers west of Del Rio, Texas show streaks and stains where moisture is held until evaporation occurs (pl. 2-B). Streaks are well displayed in Canyon de Chelly in Arizona, on Tensleep sandstone in canyons of the Wind River Mountains in Wyoming (pl. 2-C), on Park City beds north of Vernal, Utah, and in many other canyons. The red stain covering almost the entire exposure of the Redwall limestone in the Grand Canyon is a surface tension effect caused by seepage of water containing iron oxide derived from overlying redbeds and its retention on the vertical faces of the limestone.

DEVELOPMENT OF BOXHEADS

The action of surface tension in the development of overhanging rims and vertical or backsloping boxheads has been observed in a number of vertical-walled canyons in northern Caddo County, Oklahoma (Ireland, 1949). The results there are typical of other canyons in thick sandstones in many parts of the United States. The canyons are incised in a fine uniform-grained red sandstone (Whitehorse) of Permian age which is poorly consolidated and highly friable. Each trunk canyon and its tributaries begin abruptly with overhanging boxheads. Some tributaries are short re-entries with shallow boxheads, and others are merely scallops in the sidewalls of the canyons

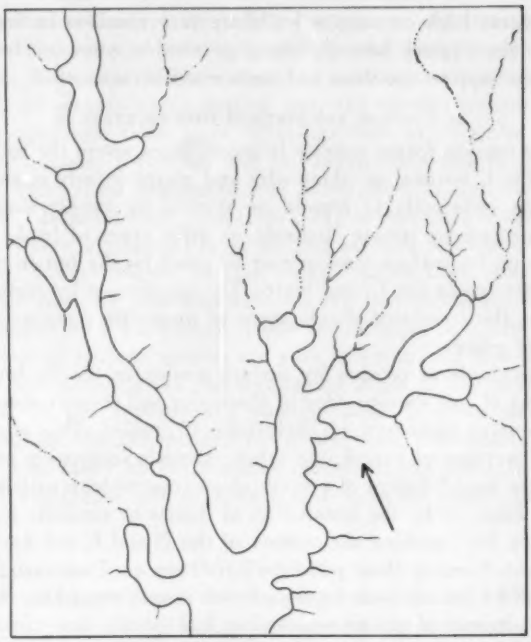


Fig. 2. Map of a portion of Kickapoo Canyon, T. 11 N., R. 11 W., Caddo County, Okla. showing many boxheads and the scallops around the rim of the canyon. Map is one mile wide.

(fig. 2). The boxheads range in height from 20 to 80 feet with overhangs of as much as 30 feet (pl. 2-D). Scour at the rims by overflow of water during wet weather has notched the boxheads and carved fluted channels in the lips. Fluted channels in some places have been incised 20 feet. Nearly all drainage channels enter the canyons as hanging valleys. Most of the canyons have runoff from springs, but runoff over the canyon rims occurs only during rains.

Each boxhead with its overhanging rim and backslope is formed chiefly by backtrickle of water held to the sloping wall by surface tension. The backtrickle removes sand particles grain by grain as the rock cement is weakened or removed. Much of the spalling and caving which enlarges the boxhead is due to seepage out of the permeable sandstone and to the root pressure of small plants attached to the moist walls. Frost probably is effective in causing some of the backslope retreat. The surge of water in the plunge basin, even though runoff from rains is rather infrequent, is responsible for some of the backslope and overhanging rim. Dark stains and streaks on the walls of the boxheads are additional evidence of the action of surface tension. Fresh scars left by falling slabs are soon streaked again, indicating the continuing action of surface tension. Streaks are also present on the sidewalls of the canyons, proving the action of surface tension there. The boxhead shown in plate 2-D has a drainage area of only a few acres and plunge pool action is very minor, yet the backslope and staining attributable to the action of surface tension is well developed.

Small caves high on canyon walls are very common in the Caddo Canyons. They are aligned laterally along permeable zones or bedding planes where seepage supplies moisture and surface tension operates.

MAGNITUDE AND DISTRIBUTION OF EFFECTS

Surface tension forces operate in most places where the soil profile contains a friable C horizon or where silty and clayey loosely consolidated sediments cohere sufficiently to remain as vertical or steeply sloping walls in roadcuts, excavations, stream channels, or other types of banks. The magnitude of erosion by surface tension may be small locally but in the aggregate it is large throughout the United States. Though erosion by surface tension is considerable, the associated development of protective painting prevents erosion in many places.

The magnitude of erosion by surface tension in the Piedmont makes it a critical area of soil erosion. One of the major soil conservation problems is the prevention or reduction of backtrickle in gullies. This may be accomplished by diversion of runoff into other channels adequately protected with vegetation, the establishment of a gentle slope covered with suitable vegetation at the gully head, or by the installation of flumes or conduits for the runoff. Any provision for covering the contact of the B and C soil horizons should reduce erosion. Some of these practices have been used successfully, but those that have failed have not been developed with proper regard for the mechanics and erosion potential of surface tension.

Painted surfaces, miniature mudflows, rill channels, muddrips, and caveheads are widely distributed in nearly all parts of the United States. In addition to the Piedmont they are very common on the Gulf and Atlantic coastal plains, in central United States, on the High Plains, in the Southwest, the Colorado Plateau, and intermontane areas. Silty and clayey Tertiary and Quaternary deposits and soils are the most susceptible to the action of surface tension, but poorly consolidated beds of other ages having the same lithology or material are affected by surface tension.

Stains and streaks are common on nearly all vertical or overhanging bluffs in either humid or arid areas.

Surface tension is a significant factor in the development of canyons with vertical or overhanging walls and boxheads with backslope. Canyons in thick upper Paleozoic and Mesozoic sandstones in New Mexico, Arizona, Utah, Colorado, and Wyoming commonly show boxheads, which are also stained and streaked. Many of these may be modified or entirely due to wind action, undermining and collapse, or separation along a joint. Many sites of ancient cliff dwellings in Mesaverde and elsewhere in the Southwest are partially or entirely due to the action of surface tension. Boxheads like those in the Caddo Canyons are found over 3,000 square miles in western Oklahoma where the Whitehorse sandstone crops out. Pleistocene deposits have filled many of them. The rate of erosion brought about by surface tension in the Caddo Canyons and in sandstone canyons elsewhere is much slower than in the Piedmont because the material is much more resistant and better consolidated.

Surface tension also operates where hot springs and some types of cold springs issue. Precipitation of calcite, silica, iron oxide, and other soluble material from the water builds terraces and small terracettes because surface tension holds or retards water flowing over the vertical terrace faces while cooling or evaporation takes place. This has been observed in Yellowstone Park, at Thermopolis, Wyoming, and in eastern Ohio and western Pennsylvania where water from abandoned coal mines deposits iron oxide.

"Hoodoo," pedestal rocks, and table rocks as described by Bryan (1925) are common where a resistant layer or slab of rock overlies a loosely consolidated bed. Water held by surface tension moves back along the undersurface of the resistant rock and removes the underlying material, leaving a pedestal capped by a hard layer. These are common in many places.

The effects of surface tension are very common and widely distributed in both arid and humid areas and throughout much of the geologic column, but especially in the Tertiary and younger beds. It seems unnecessary to elaborate here the detailed occurrence of the wide distribution. Most geologists have seen many of the features described, but many have not been aware of the significant role of surface tension in the development of those features.

CONCLUSIONS

1. Caves, gully cave-heads, overhanging boxheads and walls in canyons, painting, streaks and stains, and pedestal rocks are results of the action of surface tension commonly found in most parts of the United States.

2. Erosion by surface tension occurs where water held to a vertical or overhanging surface on loosely consolidated material moves downward carrying away particles. If an absorptive zone is present below a more resistant zone, backtrickle develops cave-heads like those in gullies in the Piedmont, and undermining and collapse follow; otherwise a backsloping boxhead or canyon wall like those in Caddo Canyons and elsewhere will develop and gradually retreat as erosion continues.

3. Painted surfaces with mudrips and mudrills do not develop in well-consolidated rocks but are common where clayey and silty poorly consolidated material occurs. The material and water, held to the face of a slope or wall by surface tension, moves downward until the water drips, evaporates, or is absorbed leaving a deposit. This aggraded surface hardens when dry and resists erosion.

4. Stains and streaks are a very common and widespread form of painting, but they have little to do with gradation, though they are evidence of the action of surface tension.

5. The effect of surface tension is slow, intermittent in some places according to the amount and rate of rainfall and seepage. The effects are small locally but are widely distributed so that the aggregate is large.

6. The action of surface tension has not been well recognized as a significant factor in the degradation of vertical and backsloping surfaces and in the aggradation of erosion-resisting painted surfaces.

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BASALT COLUMN RINDS CAUSED BY DEUTERIC ALTERATION

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ABSTRACT. A late Tertiary columnar basalt flow in southwestern Oregon is characterized by a narrow, dark rind zone around each column. This rind, which was at first attributed to chilling, has instead been found to be a zone of deuteric alteration caused by volatiles escaping toward the joint cracks. Locally the volatiles were trapped within the core of the columns and produced altered patches. The altered patches and rind zone are characterized by the presence of chlorophaeite, a hydrous mineraloid of deuteric origin. Elsewhere, the rock is fresh.

The joint cracks are believed to represent loci of reduced pressure toward which the volatile materials streamed from the interiors of the columns. Volatiles were trapped only locally in the core but uniformly in thin zones around the columns, where they proceeded to convert olivine and glassy base into chlorophaeite and, in some instances, into siderite and chlorophaeite.

Three earlier reports of joint crack control of deuteric alteration are cited. However, in all these cases the rock adjacent to the joint cracks was fresh and the rock farther away was highly altered. This condition is the *reverse* of that observed by the present writers.

INTRODUCTION

In the summer of 1949 a county road quarry in southwestern Oregon was examined. This quarry is located in the Butte Falls quadrangle in Jackson County about four miles west of Butte Falls on the road to Medford (fig. 1).

The rock is a late Tertiary (Pliocene ?) porphyritic olivine basalt flow, the "later intracanyon basalt" of Wilkinson, et al. (1941), that extends westward down the divide between McNeil Creek and Butte Creek valleys. The total observed thickness of this flow is about 40 feet, with columnar joints well developed throughout the total thickness and over an area examined of about 500 square yards. The average diameter of the columns is about two feet, and each column is surrounded by a relatively uniform black rind zone. This paper presents the results of a study of these rind zones.

DESCRIPTION OF THE BASALT

General nature.—The bulk of each column of this olivine basalt flow consists of a core zone of dark gray fresh rock completely surrounded by a narrow black altered rind zone containing the mineraloid chlorophaeite. Most columns exhibit a thin outermost yellowish-brown weathered zone. Vesicular patches occur sporadically throughout the core.

Core.—The core comprises the greatest part of the column and is characterized by clear light gray euhedral to subhedral plagioclase phenocrysts and fewer small grains of yellow olivine and dark brown augite in a black aphanitic groundmass. The phenocrysts are generally random, but locally flow structure, approximately normal to joint columns, is well developed.

Rind.—Surrounding each column is a thin black zone which, upon casual examination, appears to be an aphanitic chilled border. The zone is uniform in thickness around any given column and varies but slightly from one column to another, the average thickness being 1 cm. The contact between this zone and the core is abrupt and regular. Plate 1-A shows such a zone. Identical thin zones have been found cutting through the core in random manner, and

Fig. 1. Index map showing location of Butte Falls quadrangle.

as irregular and branching streaks. These may cut completely through the column or extend for only a short distance. Some are subparallel to the outer rind.

Close examination reveals phenocrysts of the same size and composition as those in the core, and many are seen to extend across the contact of these two zones (pl. 1-B). This zone, which is so striking in hand specimen, would go unnoticed in thin section except for its alteration products (cf. below). The plagioclase appears dark brown and glassy and shows no contrast with the dark groundmass. The mafics appear the same as in the core.

Weathered rind.—A weathered zone, varying from less than 1 mm to 5 mm occurs around the basalt columns. A thin film or coating is the most characteristic form. Weathering has softened the outer part of the rind to a yellowish-brown clay-like rock. In contrast to the unweathered rind, the phenocrysts stand out as clearly visible crystals. Except for the light brown color of the groundmass this zone presents the same appearance as the core. In thin section the rind and weathered rind are clearly demarked. All of the glass and some of the magnetite have been converted to limonite dust. The plagioclase and augite phenocrysts and the plagioclase microlites are unaltered. A few olivine grains remain. Rarely, the entire rind zone has been altered, in which cases the weathered zone is in contact with fresh core material.

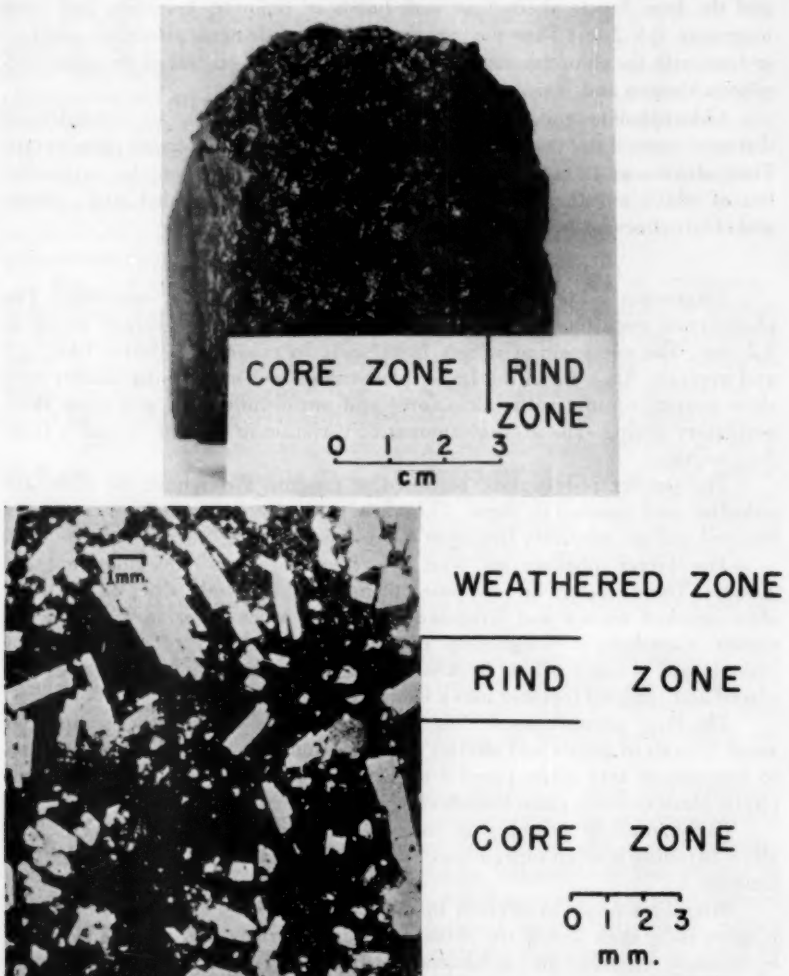
Vesicular patches.—Sparsely scattered through the core are zones of alteration containing an abundance of vesicles. These irregularly shaped areas range up to 25 cm in diameter and contain vesicles as much as 2 cm in diameter.

All vesicles are lined with chlorophaeite which in turn may be coated with siderite and calcite or natrolite. That portion of the basalt immediately

enclosing these vesicles and amygdales is very dark in appearance and resembles the rind zone.

Microscopic examination reveals alteration of the same sort that characterizes the rind zone, which is even more intense locally. The vesicle walls

PLATE 1



A. Hand specimen of basalt showing core zone and rind zone.

B. Photograph of thin section illustrating absence of chilled zone. Weathered zone at top is lighter colored. Both rind and core zones are texturally the same. Compare with A. Large clear laths are plagioclase phenocrysts; minute clear laths are plagioclase microlites. Light gray prisms are augite phenocrysts.

are lined with orange-brown and/or green chlorophaeite layers which grade into a green surface fringe of chloritic fibers. This outer chloritic fringe, which represents the breaking down of chlorophaeite, is lacking in those vesicles which are filled. The carbonate structure is radial and consists of concentric layers with the first few layers alternating with green chlorophaeite and the later layers alternating with bands of limonite, hematite, and some magnetite (pl. 2-A). Thus the carbonate phase of deuteric alteration overlaps in time with the chlorophaeite phase. This structure is typical of the spherosiderite in Oregon and Washington lavas.

Chlorophaeite and siderite permeate the host rock for considerable distances around the vesicles and completely replace many olivine phenocrysts. They also occur in patches that may have been smaller vesicles originally, but of which no other trace remains. Chloritic fibers mingled with siderite and chlorophaeite give a fine felted texture to these patches.

PETROGRAPHY OF THE BASALT

Plagioclase.—Plagioclase occurs as phenocrysts and microlites. The phenocrysts constitute 22 percent of the rock and have an average length of 2.2 mm. The composition ranges from sodic to calcic labradorite (An_{50-68}) and averages An_{63}. All of the larger phenocrysts and some of the smaller ones show normal zoning, with calcic cores and more sodic rims, and some show oscillatory zoning. The average amount of variation in a single crystal is from An_{57} to An_{63}.

The smaller phenocrysts, scattered at random throughout the rock, are euhedral and tabular in form. They are generally not zoned or complexly twinned and are relatively free from inclusions.

The larger phenocrysts tend to group together into glomerophyric clusters. These clusters are composed of individual crystals which in the main show marked zoning and irregular contacts between other members of the cluster. Cataclasis is conspicuous in these clusters. Many of the constituent crystals are sheared off or broken into smaller parts which have been displaced and jumbled together into a complex aggregate.

The large phenocrysts enclose glassy groundmass, magnetite grains, and small crystals of augite and olivine. The inclusions of glass and magnetite tend to concentrate into zones parallel to the clino-pinacoid, and in the glomerophyric clusters along grain boundaries.

Carlsbad-albite twinning is common and pericline twinning rare. The albite twinning is often superimposed upon several grains of the glomerophyric clusters.

Microlites occur in crystals up to 0.1 mm in length. Their composition is more sodic than that of the phenocrysts and averages An_{40}. Albite twinning is common, Carlsbad or Carlsbad-albite twinning is rare. For the most part the microlites are scattered haphazardly throughout the glassy base but they develop a distinct flow alignment near most phenocrysts and vesicles.

The plagioclase crystals in the core and rind zones are alike in every respect except that in the rind zone films of chlorophaeite coat the cracks and surfaces of the crystals.

Augite.—Augite is present both as phenocrysts, as much as 5 mm in length, and as minute granules in the groundmass. The phenocrysts, which appear in thin section as elongate euhedral prisms and subhedral partially resorbed crystals, comprise 5 percent of the rock. The mineral is faintly pleochroic from pale green to brownish green. Much pigeonite "molecule" is present as shown by measurements of the optic angle about Z which range from 39 to 48°. The optic plane is (010) and the maximum extinction ($Z \wedge c$) obtained was 42°. The index of refraction $N_g = 1.715$. All optical data combined indicate an Fe:Mg ratio of about 7:10. Inclusions are common in the phenocrysts and are olivine, plagioclase, grains of magnetite, and more rarely, blebs of the glassy groundmass. The large augite phenocrysts seldom cluster into glomerophyric aggregates.

Augite granules in the groundmass are less than 0.01 mm in diameter and are associated with olivine granules of about the same size. They appear predominantly brownish and are of about the same composition as the augite phenocrysts as judged by measurements of 2V.

Hypersthene.—Hypersthene occurs throughout the rock as a minor accessory and was found in about half of the 18 sections examined. It appears as elongate tabular crystals, the length ranging up to 3 mm and averaging about 1 mm. Pleochroism is distinct with X reddish orange, Y yellowish brown, and Z very light green. The 2V about X is large and appears to be about 60°, indicating an excess of Mg over Fe. There is distinct basal parting.

Olivine.—Olivine occurs as euhedral to subhedral phenocrysts, as much as 2 mm in diameter, and as small granules in the groundmass. The phenocrysts comprise about 3 percent of the rock and tend to unite with augite in cumulophyric aggregates. The olivine is clear except for a few inclusions of magnetite and glass. The grains are colorless in thin section, but pleochroic in thicker fragments from pale grayish yellow to dusky greenish yellow. They are occasionally twinned and frequently have numerous small cracks. Cleavage is distinct (010) and less distinct (100).

Dr. H. S. Yoder, Jr., very kindly made an X-ray examination of this material and found it to have $31(\pm 2)$ percent of the fayalite molecule. The following optical properties were determined: $N_m = 1.722 \pm 0.001$, $(-)$ $2V_{Na} = 80°$, $r < v$. These indicate hyalosiderite, in keeping with the X-ray determination.

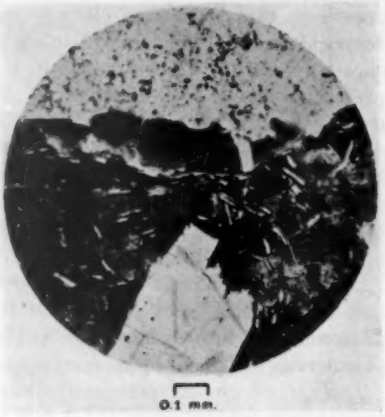
Groundmass.—The groundmass is characteristically black and comprises about 70 percent of the rock. It consists of opaque black glass and innumerable small microlites and crystallites of plagioclase. Olivine and augite granules make up about 15 percent of the groundmass. Magnetite is also present in scattered grains and in a dusty form so abundant as to make the glass opaque.

Chlorophaeite.—Chlorophaeite, the hydrous mineraloid which occurs as the alteration product in the rind zone material, was described in detail by Peacock and Fuller (1928). The optical properties of the chlorophaeite determined by the present writers correspond with those given by Peacock and Fuller.

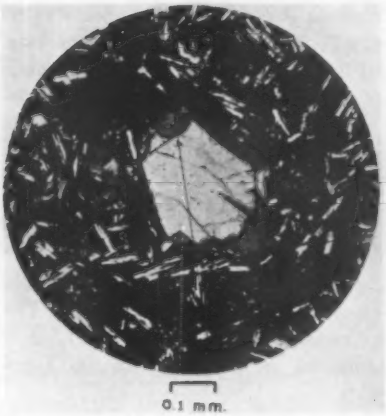
In hand specimen chlorophaeite is reddish brown. When the specimen is freshly broken the chlorophaeite, coating and partly filling some of the

PLATE 2

A.



B.



C.

D.

A. Photomicrograph of amygdale of banded siderite, chlorophaeite, magnetite and limonite. Groundmass has many patches of chlorophaeite, chlorite and siderite in felted masses (light gray irregular areas). Clear laths are plagioclase phenocrysts. Plane light.

B. Photomicrograph showing margin of vesicle lined with chlorophaeite. Outer edge of chlorophaeite crust is brown, inner portion and groundmass material are light green. Color change caused by oxidation. Large plagioclase phenocryst in lower middle; small clear laths are plagioclase microlites. Plane light.

C. Photomicrograph of olivine phenocryst partially replaced by chlorophaeite (dark gray). Surrounded by plagioclase microlites. Rind zone. Plane light.

D. Photomicrograph showing plagioclase phenocrysts with chlorophaeite coating cracks. Rind zone. Plane light.

vesicles, is green and distinctly pitch-like in luster. After a brief exposure to the air the green color changes to reddish brown.

In thin section, chlorophaeite is seen to be confined to the rind zone material and to local vesicular patches in the interior of the columns. The chlorophaeite is isotropic and is mostly reddish brown and green, and sometimes brown. The average index of refraction of a dozen grains tested is 1.51. The brownish material apparently represents an oxidized phase of the green unaltered chlorophaeite. Evidence of the color change by oxidation is shown in thin section in which a vesicle lining of the green material is coated by the reddish brown chlorophaeite (pl. 2-B). The green chlorophaeite tends to develope minute fibers commonly oriented in a spherulitic manner, but may occur as a confused mass in which the fibers are diversely oriented. These fibers, which exhibit low birefringence, are probably chlorite.

The cholorophaeite is commonly associated with olivine which it replaces along cracks and cleavages. Usually olivine is only partially altered (pl. 2-C), although all stages of alteration are found, even to complete pseudomorphic replacement. Some irregular patches of alteration occur in the groundmass and have resulted from a replacement of both the vitreous base and the groundmass olivine granules. The augite was nowhere observed to be more than incipiently replaced by chlorophaeite, although in the rind zone its cracks are commonly coated with chlorophaeite. Hypersthene was not observed in the rind zone, consequently its susceptibility to alteration is unknown. Chlorophaeite does not replace plagioclase but is found as coatings and thin films lining cracks in the plagioclase (pl. 2-D). In some cases very thin stringers or veinlets filled with chlorophaeite traverse the rind zone.

Inclusions of glassy base and olivine have been altered in some cases, while the enclosing plagioclase or augite phenocrysts are not altered even though they are traversed by chlorophaeite-filled cracks.

ORIGIN OF THE RIND ZONE

The concentration of chlorophaeite in the rind zone suggests the manner in which is was formed. Subsequent to the outpouring of the basaltic lava, the flow began to solidify at its surface by the loss of heat and gaseous material to the atmosphere. Stress, initiated by the cooling of the mass, caused centers of contraction to form near the cooling surface producing typical columnar joints. Joint cracks thus formed created loci of reduced pressure and a streaming of volatiles took place from the interior of the columns towards the outside. The basalt had cooled so that it was partially crystallized, yet fluid enough to allow a transfer of gaseous materials. That small amounts of volatiles were trapped in the columns is shown by the scattered vesicular patches; however, most volatiles probably reached the joint cracks and escaped into the air. The residual volatiles that did not quite reach the cracks were trapped in the thin zones surrounding the columns. These residual volatiles, unable to escape, attacked the olivine and glassy base and partially or completely altered them to chlorophaeite and/or siderite. The alteration appears as rims surrounding the partially altered olivine, as pseudomorphs of the completely altered

minerals, and as minute films coating and filling cracks in the plagioclase phenocrysts.

This is in agreement with Peacock (1930, p. 172) who states, "Chlorophaeite, on the other hand, according to every careful study, has been formed in doleritic rocks by the reaction of late magmatic, therefore internal, fluids, with newly crystallized ferromagnesian minerals in the rock, most commonly olivine, or sometimes perhaps, with ultrabasic vitreous residue of similar composition. The resulting gel-like product has thus a strongly basic composition corresponding roughly with that of chlorite."

SOME OTHER OCCURRENCES OF CHLOROPHAEITE ROCK

A deuteric process similar to that postulated by the present writers has been reported by Fuller (1938) in a Tertiary basalt flow that caps Steens Mountain in southeastern Oregon. Although the processes were essentially the same, Fuller indicated that the basalt columns show a marked deuteric alteration in the *central portion* of the columns and that the basalt is fresh adjacent to the major joint cracks. Fuller postulates that, as the jointing proceeded, gases escaped along the joint cracks, permitting the basalt to crystallize adjacent to the column walls before final consolidation within the columns.

In an investigation of the Iron Springs district of Utah, Mackin (1947, p. 43-47, summarized 1954) outlined a mechanism of origin of emanations similar to that of Fuller (1938). In the Three Peaks, Granite Mountain, and Iron Mountain laccoliths studied by Mackin, monzonite was completely "stewed" and deuterically altered except where tension joints penetrated into the partly consolidated crystal mesh and allowed iron-rich deuteric emanations to escape from a narrow zone of the adjacent wallrock. This escape halted the alteration and left the selvage rock relatively fresh and also lower in mafics, lower in iron content and lighter in color. He found that mineralizing emanations had been given off from the monozite only where cross joints provided channels of escape for deuteric volatiles.

Fuller has more recently (1950) described the olivine-rich zone of the Palisades diabase sill which appeared to him to be identical to the Steens Mountain flow (Fuller, 1938).

In all three instances cited above, the position of unaltered and altered zones is the *reverse* of that observed by the present writers. To the authors' knowledge no comparable examples have been described before.

SUMMARY

There are at least three processes that might have formed the altered rind of the present study: (1) weathering, (2) exomorphic alteration, and (3) endomorphic alteration. The writers have discarded weathering because of the existence of a true, vastly different weathered zone in which limonite and clay have formed, and because of the presence of completely isolated rind zone material in the fresh core. These same isolated patches prompt them to dismiss the possibility of any other exomorphic origin of the alteration and to accept the endomorphic process described above.

The most unusual feature of this basalt flow is the presence of a thin black zone surrounding each column. At first sight the rind appears to represent a thin chilled margin. However, upon examination of thin sections it was found that the two rock units are identical in texture. ***These rinds are joint-controlled zones of deuteric alteration.*** The pseudo-chilled effect can be attributed directly to the decomposition product chlorophaeite, which coats the light-colored plagioclase phenocrysts and causes them to blend in color with the black groundmass.

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U. S. GEOLOGICAL SURVEY
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REVIEWS

Le modelé glaciaire et nival; by JEAN TRICART and ANDRE CAILLEUX. Cours de Géomorphologie, pt. 2, fasc. 1, no. 2. P. iii, 408. Paris, 1953 (Centre de Documentation Universitaire).—This is an excellent and very useful summary of the physical aspects of glacial geology, and of glaciologic data useful in geologic interpretation. It is essentially a text for advanced students, beginning with the distribution of glaciers, present and past, considering glaciologic data and the geologic work of glaciers, and concluding with the effects of glaciers on crustal stability and on the level of the sea. Each chapter is followed by a selected bibliography (not unnaturally containing more European sources than American ones). The discussion is comprehensive and on controversial questions summarizes alternative points of view. The major defect of the volume is that the book itself is not well made; blurred lithoprinting of typewritten text on paper of poor quality makes the text difficult to read and detracts from the value of well-selected illustrations taken from other textbooks and from the research literature.

RICHARD FOSTER FLINT

Mountain Building; by REIN W. VAN BEMMELEN. P. xii, 177; 51 figs. The Hague, 1954 (Martinus Nijhoff N.V., 12.50 guilders).—The author of *Mountain Building*, Professor van Bemmelen of the University of Utrecht, a recognized specialist in tectonics, is well known, especially to European geologists. He has had a long and varied field experience in many parts of Europe, the British Isles, Iceland, and Indonesia, and has published extensively on theoretical aspects of orogenesis. *Mountain Building*, his latest publication, is divided into two sections. Part I, 34 pages in length, treats of theoretical considerations in the process of mountain building, and Part II is a summary of the general geology of Indonesia as well as an interpretation of the orogenic evolution of the Earth's crust in that region.

In Part I Professor van Bemmelen reviews most rapidly the more important classical and current theories of mountain building as well as concepts and problems to be resolved by a satisfactory theory. He then presents his own theory in some detail, applying it to the Tello-Betic mountain system of the West Mediterranean area. Problems and concepts which the author outlines and evaluates are whether the source of endogenic energy is a thermal or chemical process or both; the relative importance of tangential compression of the crust as the main cause of orogenesis as opposed to primary vertical forces and secondary spreading by gravity; and finally the relative importance of general compression in opposition to local mutual compensation of compression and extension. About half of the first part of the book is devoted to a discussion and evaluation of two groups of current theories of mountain building. First he presents that class of theories favored by the majority of contemporary geologists and geophysicists. These advocate tangential compression as the basic mechanism of orogenesis with a thermal source of energy. He refers to them as unicausal-mobilistic theories. In the latter half of Part I the author presents in more detail his own theory based on the geochemical concept of crustal evolution and develops the idea that orogeny is the result of geochemical processes at depth. The author refers to

the class of theories to which his own belongs as the bicausal-fixistic type. He concludes this treatment of the principles of mountain building by applying his theory to data from the Tello-Betic system. These discussions in Part I are intended to introduce the reader to the theoretical concepts involved in the discussion of the geological evolution of Indonesia treated in the second part. One would have hoped in a book such as *Mountain Building* to find a more comprehensive and more exhaustive survey of the principles of orogenesis than is found in this 34-page treatment of the subject.

Part II contains a most interesting presentation in summary form of the major features of the orogenic history of Indonesia as known at present. This study is an extension of the survey of the geologic literature of Indonesia which van Bemmelen made several years ago at the request of the Netherlands East Indies Government. Here the author reviews in 133 pages the stratigraphy of Indonesia, the active and paleo-volcanic history, gravitational tectogenesis as exemplified by that region, the data of geophysics, and finally the geological evolution of the Sunda region which he describes period by period. The stratigraphy is treated systematically and the general relationships are emphasized. Inasmuch as the discussion is limited to 17 pages, only the most important features are reviewed. An extended treatment of 31 pages is given to the subject of volcanism, and this chapter is very informative. Van Bemmelen develops here the distribution of volcanism in time and space and brings out the relation of magmatic provinces to various structural features of the region. The final chapter on the geological evolution of the Sunda region from the Devonian to the present is a very fine 40-page treatment of the orogenic history.

The topics discussed in this second section of the book though treated briefly are covered adequately. Each chapter with the exception of that on geophysics contains information on Indonesia not as easily obtained nor as well presented elsewhere in English. The fact that the region under discussion is one of extremely high negative gravity anomalies and is one of the most orogenically active areas of the Earth makes the information here presented so concisely most valuable and extremely interesting for students of tectonics and of advanced general geology. The specialist who finds that the information contained in this book is not detailed enough may consult in addition the previously written two-volume synthesis of the general geology of Indonesia by the same author and other references cited in the text. The specialist in tectonics may not fully agree with interpretations of some of the data but he is certain to find this part of *Mountain Building* extremely interesting and most stimulating.

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